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CONTENTS

	Pages
Tertiary geology of a portion of the Central Cascade Mountains, Washington. By Robert J. Foster	99–126
Criteria for recognition of laharic breccias, Southern Cascade Mountains, Washington. By Richard V. Fisher	127–132
Intravalley variation in slope angles related to microclimate and erosional environment. By Mark A. Melton	133–144
Carbonate replacement of detrital crystalline silicate minerals as a source of authigenic silica in sedimentary rocks. By Theodore R. Walker	145–152
Transcurrent faulting and volcanism in Owens Valley, California. By L. C. Pakiser	153–160
Late Pleistocene glaciation of eastern Newfoundland. By Stuart E. Jenness	161–180
Thrust faulting and chaos structure, Silurian Hills, San Bernardino County, California. By Donald H. Kupfer	181–214
Short Notes	
Silurian fish fossils in the Salina basin. By W. P. Leutze	215–218
Potassium-argon ages in the Precambrian basement of Colorado. By C. E. Giffin and J. L. Kulp	219–222
U-Pb age of the Chattanooga shale. By J. C. Cobb and J. L. Kulp	223–224
Potassium-argon ages in northern Manitoba. By J. M. Moore, Jr., S. R. Hart, C. C. Barnett, and P. M. Hurley	225–230
Apparent-dip computer. By Lowell R. Satin	231–234

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High-Temperature Alpine-Type Peridotite from Venezuela. By D. B. MacKenzie

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Role of Hydroxyl Orientation in Mica Alteration. By William A. Bassett

Agua Blanca Fault—a Major Transverse Structure of Northern Baja California, Mexico. By C. R. Allen, L. T. Silver, and F. G. Stehli

Diastrophism and Mountain Building. By Marland P. Billings

Age of the Granitic Rocks of Nova Scotia. By H. W. Fairbairn, P. M. Hurley, W. H. Pinson, Jr., and R. F. Cormier

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TERTIARY GEOLOGY OF A PORTION OF THE CENTRAL CASCADE MOUNTAINS, WASHINGTON

BY ROBERT J. FOSTER

ABSTRACT

The Central Cascade Mountains of Washington are composed largely of Tertiary continental and volcanic rocks. They lie between older metamorphic rocks to the north, Tertiary marine rocks to the west, and younger volcanic rocks to the south and east. An eastward-dipping reverse fault through Lake Kachess separates two Tertiary basins of the same general age. This fault cuts Eocene rocks and is older than the Snoqualmie granodiorite.

East of the fault the structures are open. The Swauk arkose unconformably covers pre-Tertiary peridotite and Easton schist. The Swauk is folded along northwest trends that become more east-west south of Mount Stuart. This change in trend and the local origin of some of the Swauk suggest that the Mount Stuart block was high during early Tertiary time. Locally the Silver Pass volcanic rocks overlie the Swauk. The conformable sequence of Teanaway basalt-Roslyn arkose lies unconformably above the Swauk and is deformed into a broad basin. The nearly horizontal Yakima basalt unconformably overlies the older rocks.

West of the Kachess fault, the structures are more complex. The oldest rocks are limy hornfels and marble of the Denny formation. They are overlain apparently unconformably by the sedimentary rocks of the Guye formation. Unconformably above the Guye is the extrusive Mount Catherine rhyolite; that is overlain by the tightly folded sedimentary rocks and basalt of the Naches formation. The mildly deformed Keechelus andesite overlies the Naches unconformably. The Snoqualmie granodiorite intrudes all the units in the western area.

Fossils are rare. A few vertebrate remains indicate that the Roslyn is probably middle or upper Eocene. Fossil leaves suggest that the Swauk and Guye formations are Paleocene or Eocene and the Naches is Eocene.

CONTENTS

TEXT

	Page
Introduction	100
Geologic setting	100
Location	100
Field study	100
Acknowledgments	101
Stratigraphy	101
General statement and summary	101
Eastern stratigraphic section	102
Easton schist	102
Marble and volcanic rocks west of Cle Elum Lake	102
Peridotite and older rocks south of Mt. Stuart	102
Swauk formation	103
Silver Pass volcanic rocks	105
Teanaway basalt	107
Dike swarm and diabase associated with Teanaway basalt	108
Roslyn arkose	108
Post-Roslyn rhyolite	110
Yakima basalt	110
Howson andesite	110
Western stratigraphic section	111
Denny formation	111
Guye formation (restricted)	111
Mt. Catherine rhyolite	114
Naches formation	114
Keechelus andesite	118
Snoqualmie granodiorite	121
Structure	123
Relationship between the two sections	123
Kachess fault	123
References cited	124

ILLUSTRATIONS

Figure	Page
1. Index map of Washington	100
2. Teanaway dikes	109
3. Rampart Ridge, type area of Keechelus formation	120

Plate	Following page
1. Geologic map and cross sections, Central Cascades, Washington	126

TABLE

Table	Page
1. Diagrammatic tentative correlation chart of the eastern and western stratigraphic columns in the Central Cascades, Washington	101

Introduction

Geologic Setting

The Central Cascade Mountains of Washington are composed of folded and faulted Tertiary volcanic and sedimentary rocks. The pre-Tertiary metamorphic basement is well exposed in the Northern Cascades. East of the Central Cascades, in the Columbia Plateau, only the Miocene Columbia River (Yakima) basalt is exposed; to the south, Oligocene (?) Keechelus volcanic rocks cover most of the older rocks. To the west are Tertiary marine rocks of the same general age as the continental sedimentary and volcanic rocks discussed herein.

Pioneer work in the Cascade Mountains was done in this area, and the relationships have been used to date the history of much of the range. I. C. Russell, G. O. Smith, F. C. Calkins, and Bailey Willis did reconnaissance mapping of the Central Cascades around the turn of the century; the area has received little attention since then because of the rugged terrain, harsh climate, thick vegetation, and lack of economic incentive. Excellent though the early work was, it requires revision.

Location

This area lies near the geographical center of Washington (Fig. 1). It extends from the Cascade crest at Snoqualmie Pass 40 miles eastward to Table Mountain, the edge of the Columbia Plateau. The southern boundary is the Yakima River; the area lies entirely within the northern parts of the Mt. Stuart and Snoqualmie folios of the Geologic Atlas of the United States (G. O. Smith, 1904; Smith and Calkins, 1906). The Sunset Highway, U. S. 10, crosses the Cascades at Snoqualmie Pass and parallels the Yakima River along the southern boundary of the area.

FIGURE 1.—INDEX MAP OF WASHINGTON

Black indicates area mapped in this study. Outlined are the Snoqualmie quadrangle to the west and the Mt. Stuart quadrangle to the east.

Climate, vegetation, topography, culture, and rock exposures vary greatly within the area. Yearly rainfall ranges from about 10 inches at Ellensburg, 15 miles south of the eastern edge of the area, to approximately 100 inches at Snoqualmie Pass. Elevation ranges from approximately 2000 feet in the Yakima Valley to 6400 feet at Jolly Mountain.

Field Study

The author spent approximately 8 months in the field during the summers of 1953 through 1957. Because of the differences in vegetation and exposure, the accuracy of mapping varies considerably from place to place. Critical and well-exposed areas were mapped on aerial photographs. Most of the area is tree-covered, however, and accurate location was difficult; covered areas were mapped on enlarged copies of U.S.G.S. topographic maps. Because of this, the topographic maps on a scale of 1/125,000, the same ones used in the earlier folios, are used as the base for Plate 1. The remapping

in the Snoqualmie Pass quadrangle shows many differences from the earlier folios, but the simpler structures in the Mt. Stuart quadrangle are essentially the same as in the folio with only the addition of the Swauk folds.

Acknowledgments

Many people advised and assisted the author. This study is the outgrowth of an earlier study suggested by Dr. Howard A. Coombs, who has helped and encouraged the author in every possible way throughout both studies. Financial assistance was provided by the Geology Department of the University of Washington through a teaching assistantship from 1955 to 1957, and field expenses for the season of 1956 were paid by the Humble Oil and Refining Company. Martin L. Stout, Ross C. Ellis, and Richard M. Pratt, who mapped surrounding areas, placed their unpublished work at the author's disposal; their discussions and criticism, both at home and in the field, have contributed much to the present work.

Dr. J. A. Aurèle La Rocque of The Ohio State University studied the fresh-water mollusks, and Dr. Roland W. Brown of the U. S. National Museum identified the leaf collections. Profs. H. A. Coombs and G. E. Goodspeed loaned the author thin sections, and they and Profs. Peter Misch and J. Hoover Mackin gave freely of their time and advice on a wide variety of problems.

Stratigraphy

General Statement and Summary

The present study is based on reconnaissance mapping which is more detailed than that of earlier workers. The necessity for remapping should not be construed as a criticism of the early workers, who, although working in a pioneer area, established many of the main relationships.

The most significant result of this remapping is the recognition of two distinct Tertiary stratigraphic sections separated by a north-south fault through Kachess Lake. Tentative ages and correlations of the two sections are shown in Table 1.

East of the fault, the structures are open, and the section consists of the well-known formations of the eastern Cascade Mountains. The pre-Tertiary basement rocks are peridotite and Easton schist, both of which are covered unconformably by the arkosic Swauk formation. Locally the Swauk is covered, apparently unconformably, by andesitic rocks herein named the Silver Pass volcanic rocks. Unconformably above these volcanic rocks is the Teanaway basalt, overlain concordantly by the arkosic Roslyn formation. The Yakima basalt, the youngest unit in this section, lies unconformably on the older rocks.

Table 1.—Diagrammatic Tentative Correlation Chart of the Eastern and Western Stratigraphic Columns in the Central Cascades, Washington

Age assignments are discussed in the text under each formation. Not to scale

Age	Western Column	Eastern Column	Rock Type
Pliocene		Yakima	Basalt
Miocene	Keechelus		Andesite
Oligocene			
Eocene	Naches	Roslyn Teanaway	Arkose and basalt
	Mt. Catherine	Silver Pass	Volcanic rocks
	Guye	Swauk	Sediments
Paleocene			
Cretaceous			

West of the Kachess fault, the formations are less well known, and the structures are more complex and are in part obscured by metamorphic effects owing to the emplacement of the Snoqualmie granodiorite. In the western area, most of the formations have been named or redefined by the author. Here the oldest rocks are the pre-Tertiary marble and limy hornfels of the Denny formation. These are unconformably overlain by the sedimentary rocks of the Guye formation as restricted by the writer. Unconformably above the Guye formation is the Mt. Catherine rhyolite, which is overlain

by the Naches formation. The youngest unit is the Keechelus andesite, which unconformably covers all the older rocks. This entire western section is intruded by the Snoqualmie granodiorite.

Eastern Stratigraphic Section

Easton schist.—The pre-Tertiary basement of this area is composed of Easton schist west of the Cle Elum Valley; east of Cle Elum Valley the basement comprises peridotite and metamorphic rocks of the Mt. Stuart (or Wenatchee Mountains) block.

The Easton schist was named by G. O. Smith (1903, p. 14), who later (Smith, 1903; Smith and Calkins, 1906) expanded his description. In the area under discussion, it consists of blue amphibole schist and apparently interbedded greenschist and phyllite. It is apparently the southern end of an extensive but discontinuous unit that has been traced from near the Canadian border to just north of the Snoqualmie quadrangle (Misch, 1952; B. H. Bryant, 1955, Unpub. PhD thesis, Univ. of Washington; J. A. Vance, 1957, Unpub. PhD thesis, Univ. of Washington).

Easton schist is exposed in the cores of two anticlines between Kachess and Cle Elum Lakes and near the town of Easton.

The details of the complex internal structure of the schists were not determined because of the limited time available. However, lower-grade metamorphic marble and volcanic rocks were found (Pl. 1), apparently in fault contact with the Easton schist. For the most part, they are confined to the ridge south of French Cabin Creek and the ridge between Knox Creek and French Cabin Creek. The larger of these occurrences was described by Smith and Calkins (1906, p. 2).

AGE: The Easton schist is pre-Swauk and therefore pre-Tertiary. W. S. Smith (1915; 1916) reported Ordovician fossils north of Snoqualmie quadrangle in less metamorphosed rocks and used these to date the Easton schist as pre-Ordovician. Recent workers have not found Ordovician fossils, and the original Ordovician collection has been lost (W. R. Danner, 1957, unpub. PhD thesis, Univ. of Washington).

DESCRIPTION: Most of the Easton schist in the area mapped is blue amphibole schist. In hand specimen, these rocks are very fine-grained, are blue or greenish blue, and have a silky luster. They are 40–60 per cent epidote, 10–35 per cent blue amphibole, up to 20 per cent albite, 5–10 per cent chlorite, and have minor quartz. Of five blue amphibole schists studied, two contained crossite. Some of these amphiboles are zoned with crossite or glaucophane cores and green, apparently sodic actinolite, rims.

Greenschist appears to be less abundant than blue amphibole schist in the area mapped. These rocks are fine-grained and similar in appearance, except for color, to the blue amphibole schist. The average composition is 20 per cent epidote, 35 per cent actinolite, 15 per cent chlorite, 25 per cent quartz, and 5 per cent albite.

The phyllites are fine-grained, medium-gray rocks in which quartz and graphite can be distinguished in hand specimen. The weathered surface is commonly iron-stained. Pods and veins of quartz are common and many contain minor amounts of untwinned albite, sericite, chlorite, and graphite. A few contain stilpnomelane.

Marble and volcanic rocks west of Cle Elum Lake.—The occurrence and relationships of these rocks were described with the Easton schist.

DESCRIPTION: The marble from the ridge south of French Cabin Creek is medium-gray, medium-grained, and slightly foliated with brown bands emphasizing the foliation. It is composed mostly of large and small grains of sheared calcite and irregular isolated grains of quartz and cloudy plagioclase, both of which have wavy extinction.

The calc-phyllites on the ridge between Knox Creek and French Cabin Creek are fine-grained, light-green, schistose rocks with calcite pods and lenses. In thin section, they are seen to be composed of chlorite, quartz, carbonate, and minor epidote.

The volcanic rock on the ridge south of French Cabin Creek is statically altered tuff and lapilli-tuff. The matrix is green or purple, and the angular fragments are different shades of these same colors. In thin section, only chlorite, quartz, cloudy plagioclase, magnetite, and hematite can be distinguished. Although the rock fragments in these tuffs are altered, several different lithologic types can be distinguished. Some of the fragments are of pilotaxitic lava with the feldspars still discernible, and others have quartz grains in a chloritic matrix. There are a few small, late, quartz veinlets in these rocks.

Peridotite and older rocks south of Mt. Stuart.—The peridotite forms a crude crescent-shaped outcrop that is the southern end of the uplifted Mt. Stuart block. It has been described by Russell (1899, p. 109), G. O. Smith (1903; 1904), and Smith and Calkins (1906).

In the present study, the contact between the peridotite and overlying Swauk was the map boundary. The peridotite is more or less completely serpentinized. Smith (1904, p. 4) noted reddish brown and green as the most common

colors, and erosional forms ranging from steep slopes of huge talus boulders to low rounded hills. In hand specimen, this rock is commonly mottled green and has a waxy luster. Under the microscope, remnants of the original olivine are seen along with some amphibole, pyroxene, and magnetite. The peridotite is pre-Swauk, and the basal Swauk rocks are locally composed of peridotite detritus.

Swauk formation.—The oldest Tertiary unit in the eastern section is the Swauk formation, composed of arkose, shale, and conglomerate. These rocks were first described by I. C. Russell (1893, p. 20), who referred them to the "Kittitas system." In that paper, Russell did not distinguish between what are now the Swauk and Roslyn formations. However, 6 years later, he recognized and named both formations and alluded to a possible subdivision within the Swauk. He stated (1899, p. 118),

> "The rocks included in this formation [Swauk] present two quite distinct phases, which led me to divide them into two systems, one termed the Camas sandstone and the other the Wenatche sandstone; subsequently, however, these terranes were studied, in part in considerable detail, by Messrs. Willis and Smith and . . . were found by them to be deposits of a single Tertiary Lake or estuary; therefore, the name Swauk sandstone was given to the entire formation after the Swauk mining district where it occurs."

The Swauk is now recognized as one of the most widespread formations in the Cascades. G. O. Smith (1904; Smith and Calkins, 1906) mapped the Swauk formation in the area under discussion. The Swauk formation was traced far north by Russell (1899) and west by W. S. Smith (1916).

After remapping the northwest portion of the Mt. Stuart quadrangle, Frank Alexander (1956, Unpub. MS thesis, Univ. of Washington) favored subdivision of the Swauk into two formations separated by a fault. Fossil-leaf localities in nearby areas, apparently on strike with his subdivisions, have the same flora (*cf.* A. C. Waters, 1930, Unpub. PhD thesis, Yale Univ.); however, these plants had long ranges. Chappell (1936, Unpub. PhD thesis, Univ. of Washington, p. 93) mapped an unconformity in what he called Swauk in the Wenatchee quadrangle east of Alexander's area. The importance of this unconformity is unknown, and it has not been reported elsewhere.

Lupher (1944, p. 7, 8) discussed variations in the particle size and the induration of the Swauk in the area of the present study. G. O. Smith was aware of these differences in the Swauk formation but minimized their importance. He stated (1904, p. 5),

> "The general character of the Swauk sandstone is that of an arkose. It is usually plainly bedded, and interstratified with shaly and conglomerate beds. In color the sandstone is gray, and light grains of feldspar and quartz and dark flakes of mica may be noticed. In the eastern part of the area, especially along Mission Creek and its tributaries, the Swauk formation shows a notable change in character. The shale and conglomerate become insignificant in amount, and the sandstone is lighter colored and plainly more purely quartzose. This phase of the sandstone is massive and less plainly bedded, and doubtless represents different conditions of sedimentation. . . ."

Smith used the description of the rocks along Swauk Creek in both the Mt. Stuart and Snoqualmie folios, even though the indurated rocks in the Snoqualmie quadrangle bear little resemblance to the nonindurated rocks to the east.

OCCURRENCE AND RELATIONSHIPS: The Swauk crops out in a broad band along the northern border of the map area. Throughout much of this area, the Swauk is intruded by the Teanaway dike swarm. Smaller areas of Swauk are also found between Kachess and Cle Elum lakes.

The Swauk exposures in the area of the dike swarms are remarkably sparse. In the dike swarm, the Swauk forms thin septa, many of which are only a few feet wide and pinch out in short distances. These septa are commonly widely separated and are less resistant than the dikes and so are covered with basalt talus. The abundance of dike basalt causes difficulty in location of the contact between the Swauk sedimentary rocks and the Teanaway basalt flows. The Swauk septa are also remarkably undeformed. The attitude of the beds in these septa is constant over wide areas, and the Swauk folds were mapped by study of these septa.

At least part of the basal Swauk is of local origin. G. O. Smith (1904, p. 5) noted that ". . .the coarsest phases of the basal conglomerates are very local in their occurrence, and in their composition often bear a definite relation to the underlying formation."

In the western part of the area, the Swauk is folded along northwesterly trends and the Easton schist is exposed on the eroded anticlines. South of the Wenatchee Mountains, folding is stronger, trends nearly east-west, and only the Swauk is exposed.

Where the Swauk unconformably overlies peridotite, lateritic iron ores have formed.

These ores were first described by Kimball (1898); Smith and Willis (1901) described them more completely. There have been many attempts to exploit these deposits, but their small size and irregular form have prevented them from being of commercial value.

Most authors have considered these laterites the basal Swauk; however, Lupher (1944, p. 7) proposed the name Cle Elum formation for these deposits. The situation is further complicated by the fact that the iron ores show definite evidence of limited transportation, although their local source is recognized. Lupher (1944) and Lamey and Hotz (1952) give a discussion and bibliography.

In most of this area, the Swauk is overlain unconformably by the Teanaway basalt and, locally between Kachess and Cle Elum lakes, by the Silver Pass andesite. The unconformity and the pre-Teanaway folding of the Swauk were noted by G. O. Smith (1904, p. 5). Russell had reported (1899, p. 122) that the Swauk dipped homoclinally southward in the area south of Mt. Stuart, but that it was folded in other areas.

Minor faulting and brecciated zones near the base of the Swauk (Russell, 1899, p. 120) indicate that it has undergone more than one period of diastrophism. Lupher (1944, p. 27–28) reported 2000 feet of offset on one fault and less than 200 feet of offset on many others.

AGE AND CORRELATION: The age of the Swauk formation has been much debated. F. H. Knowlton (*in* G. O. Smith, 1904, p. 5), on the basis of about 25 new species of fossil leaves collected on Swauk Creek near Liberty, reported ". . . a more or less close resemblance to certain Laramie, Denver, and Fort Union species [exists], and on this rather insecure basis, it is assumed that the age should be regarded as Eocene." The term Paleocene had not yet come into use, and it seems clear from Knowlton's reference to what are now considered Cretaceous and Paleocene formations that he regarded the Swauk as being near the Cretaceous-Tertiary boundary, not Eocene as we use the term today. Many, however, have continued to call the Swauk Eocene on the basis of Knowlton's report.

The Paleocene or Late Cretaceous age based on paleobotany is in accord with the stratigraphic relationships. The Swauk unconformably overlies a metamorphic basement of unknown age. Unconformably above the Swauk is the Teanaway basalt, and concordantly above the Teanaway is the Roslyn arkose that contains middle or upper Eocene fossils (Wheeler, 1955, p. 1668). Thus, the evidence suggests a Paleocene or Late Cretaceous age for the Swauk, but more secure dating must await future fossil discoveries.

CONDITIONS OF DEPOSITION: Almost all the early geologic reports on the western United States describe the nonmarine sedimentary rocks as lacustrine deposits. This dogma began in the Rocky Mountain area and was carried into the Cascades by the early workers. The descriptions by G. O. Smith and Russell are quite accurate, so that the careful reader may become suspicious of the alleged lacustrine origin of these rocks. Russell (1899, p. 125) described "thin bands of well-worn pebbles" in the Roslyn formation. G. O. Smith (1904, p. 7) described the same rocks as follows:

"With the sandstone occur shales, both fine-grained clay shales and the coarser arenaceous phase. As a rule, the stratification of these rocks is not strongly marked, and in some localities irregularities of bedding can be seen and local unconformities detected. Conglomeratic beds are not common, pebble bands in the sandstone being the coarsest material usually found in this formation."

The meaning of these ubiquitous fluviatile features in the Swauk formation was first noted by W. S. Smith (1916, p. 565), and they have been reported by A. C. Waters (1930, Unpub. PhD thesis, Yale Univ.), W. M. Chappell (1936, Unpub. PhD thesis, Univ. of Washington), C. L. Willis (1950, Unpub. PhD thesis, Univ. of Washington), and others. In the area under discussion, fluviatile features such as cross-bedding, alternating coarse- and fine-grained layers, gravel lenses, local unconformities, cut and fill structure, and terrestial (leaf) fossils are common.

DESCRIPTION: The Swauk formation is composed predominantly of arkose, with some shale and minor conglomerate. Rock types are distributed irregularly. The beds range from a few inches to more than 100 feet thick, and grain size changes abruptly both laterally and vertically, so that it is impossible to describe a section applicable to more than a small area. Most of the Swauk is massive, but cross-bedding is not uncommon.

The Swauk is at least 4000 feet thick in this area and may be much thicker. The dike swarms and the lack of key beds make it impossible to estimate the thickness more closely. Smith described a basal conglomerate, but more detailed work has shown that in most places, the basal Swauk is shale or siltstone that grades downward into the iron-rich laterite (Lamey and Hotz, 1952, p. 32). Above these

basal rocks is arkose with some interbedded dark shale. Conglomerate occurs intermittently throughout the section, some lying directly on the basement rocks.

The arkose is light gray on the fresh surface and weathers to brown or brown gray. It is medium- to coarse-grained, and quartz, feldspar, and biotite are visible in hand specimen.

In thin section, the arkose is seen to be composed of angular equant grains, some of which show pressure solution. The principal minerals are quartr, 40–50 per cent; feldspar, 20–40 per cent; and biotite, up to 10 per cent; much of the biotite is altered to chlorite and magnetite. Most of the feldspar is twinned plagioclase, much of which is turbid, and is primarily oligoclase but ranges from albite to andesine. The minor constituents are muscovite, orthoclase, garnet, magnetite, chlorite, apatite, sphene, and in some specimens, rock fragments, especially chert, and quartzite. This arkose is fairly well sorted, and there is little matrix. The cement is silica in most sections, but calcite is not uncommon. Near the dikes, the alteration of feldspar and biotite is pronounced, and some quartz grains are sheared and healed with quartz and calcite.

The shale is dark gray to black and weathers to medium brown. At some horizons, leaf fossils and carbonaceous material are common. Many of these rocks are fine-grained siltstones or mudstones composed of the same minerals as the sandstone. The beds range from a few inches to many feet in thickness and are interbedded with the sandstone, commonly with sharp contacts, but some of the shales are sandy.

The conglomerate is scattered through the section in thick and thin beds. Much of the conglomerate is composed of rounded cobbles, ranging from 2–6 inches in diameter, in an arkosic matrix. In the area, the cobbles are predominantly quartzite, but light-colored gneiss, vein quartz, and granitic rock types are common.

Silver Pass volcanic rocks.—The name Silver Pass volcanic rocks is proposed for the unit composed mainly of andesite and more felsic volcanic rocks, including flows, tuff, and breccia, that is exposed near Silver Pass. Smith and Calkins (1906, p. 5) included these rocks in their Kachess rhyolite, although they did comment on the difference between these rocks and the Kachess rhyolite. They observed

> "A thicker bed of andesite constitutes the base of the formation west of the head of Clealum Lake and is poorly exposed on the eastern side of Kachess Lake. This rock is of light greenish-gray color and is not conspicuously different from the rhyolite."

They named the Kachess rhyolite "... for Kachess Lake, on whose northeast side it is well exposed." Thus the Silver Pass volcanic rocks include the "type area" for Smith and Calkins' Kachess rhyolite; however, mapping for the present study has shown that rhyolites of many different ages were included in the Kachess rhyolite by Smith and Calkins. Smith and Calkins were aware of these inconsistencies, and they described the relationships of the Kachess as peculiar, noting that it was interbedded with the Swauk, Teanaway, and Naches formations and occurred between the Naches and Teanaway formations. These relationships would require the extrusion of the Kachess to have begun in Swauk time, to have extended across the unconformity between Swauk and Teanaway, and to have continued until post-Teanaway time. In the area mapped, the rocks called Kachess by Smith and Calkins can be separated into two groups, presumably of different ages, the Silver Pass volcanic rocks of post-Swauk and pre-Teanaway age, and the rhyolite within the Naches formation.

OCCURRENCE AND RELATIONSHIPS: Silver Pass, the type area, is between Silver Creek and French Cabin Creek, both of which drain the region between Kachess and Cle Elum lakes. At Silver Pass, the volcanic rocks unconformably overlie Easton schist, but southwest of here, on the west ridge of French Cabin Mountain, they overlie Swauk formation. Near Silver Pass these rocks are intruded by basaltic dikes, presumably the Teanaway dike swarm. Farther south, on the ridge between Kachess Lake and Silver Creek, Teanaway basalt overlies Silver Pass rocks, apparently unconformably. Thus the Silver Pass volcanic rocks are younger than Swauk and Easton and older than Teanaway.

The Silver Pass volcanic rocks form rugged pinnacles and spires in the French Cabin Mountain-Silver Pass area. Near timber line, the andesites weather light gray, and from a distance, the red-brown-weathering basalt dikes are striking. The unconformable contact between the Silver Pass and the Swauk can be seen on the ridge trending more or less east from French Cabin Mountain. On the south-trending ridge near this point is the contact with the Teanaway. This contact occurs in a gully between two of the minor summits of the south-trending ridge. This is the only place where the Teanaway-Silver Pass contact can be located; elsewhere it is mapped on the basis of float and isolated outcrops. The two localities are shown by solid-line contacts on Plate 1. The nature of the Teanaway-Silver Pass contact in the area south of French Cabin Mountain is a problem. As mentioned, it is exposed at only one place, and even there the relationships are not clear.

Dikes occur in that area, and because the rocks mapped as Teanaway are aphanitic to coarse-grained basalt, the contact may be intrusive. Teanaway basalt flows occur south and west of this area, so this exposure is tentatively mapped as Teanaway basalt formation. Southwest of this point across Silver Creek Valley northeast-dipping Teanaway rocks crop out. The nature of the contact across Silver Creek Valley is not known because there are no exposures, but it may be unconformable, or it may be a fault.

The Silver Pass volcanic rocks have been recognized only in the area described herein, its eastern extension near Nelson, and in a small occurrence east of Thomas Mountain. Elsewhere the Teanaway basalt overlies the Swauk directly, and the Silver Pass volcanic rocks, if ever present, have been removed. A small body of similar andesite crops out near the head of Cle Elum Lake and has provisionally been included with the Silver Pass volcanic rocks. Upslope from this latter area is an andesite dike of unknown age.

AGE AND CORRELATION: No fossils have been found in the Silver Pass volcanic rocks. They lie unconformably on the Swauk that is perhaps Late Cretaceous or Paleocene. Unconformably above the Silver Pass volcanic rocks is the Teanaway basalt, and concordantly above the Teanaway is the Roslyn arkose of probable middle or late Eocene age (Wheeler, 1955, p. 1668). Therefore, the Silver Pass volcanic rocks may be Paleocene or Eocene, but the discovery of fossils may alter this interpretation.

The correlation of the Silver Pass volcanic rocks presents many problems. Smith (1904) and Smith and Calkins (1906) recognized the Taneum andesite in the area south of the Yakima River. It is more or less on strike with the Silver Pass volcanic rocks and is similar lithologically. It is described as lying between the Manastash formation and the Yakima basalt. M. L. Stout has restudied this area recently. While mapping, Stout (1957, Unpub. MS thesis, Univ. of Washington, p. 25–53, 72–84, 100) believed that the Manastash might be the equivalent of the Swauk and that it was intruded by dikes similar to the Teanaway dikes. As a result of later work, covering a larger area, Stout (1959, Unpub. PhD thesis, Univ. of Washington) now believes that the Manastash and at least part of the Taneum of Smith are parts of the more widespread Naches formation. Since both Stout and the present author believe that the Naches and the Teanaway basalt are at least in part correlative, the relationship between the Silver Pass and the Teanaway may be more complex than the present mapping indicates; however mapping for this paper shows clearly that at least in the area discussed herein the Teanaway basalt is unconformable on the Silver Pass volcanic rocks.

DESCRIPTION: The Silver Pass volcanic rocks form a thick unit composed of lava, pyroclastic rocks, and volcanic sedimentary rocks. (The nomenclature of Wentworth and Williams, 1932, is used throughout this paper.) The unit is heterogeneous and characterized by abrupt horizontal and vertical changes in rock type.

Andesite comprises the majority of the lavas and most of the rock chips in the fragmental rocks. All the rocks in the Silver Pass formation are altered to some degree. In hand specimen these rocks range from light to dark green, and some are dark purple or gray. The weathered outcrop is commonly brown or light gray. Plagioclase phenocrysts and, in some rocks, small white or green amygdules are visible.

The andesites have plagioclase phenocrysts in a pilotaxitic groundmass, and many contain a few pyroxene phenocrysts. The plagioclase is euhedral to subhedral andesine, and many of these phenocrysts are zoned with slightly more calcic cores. These feldspars are cloudy with inclusions of kaolin, sericite, and chlorite. The scattered pyroxene phenocrysts are subhedral to anhedral unaltered augite (2V of 35°). Scattered masses of chloritic material or antigorite appear to be pseudomorphs after ferromagnesian minerals. The groundmass is composed in most cases of plagioclase microlites, chloritic material, and opaque minerals. The microlites are oligoclase, about An_{27}. Sphene is present in most of these rocks as inclusions in the plagioclase phenocrysts and as small scattered grains in the groundmass.

The rhyolite and dacite are commonly light green or light pink, but some are light to medium brown. They weather to brown or red brown. In a few of these rocks, clear quartz crystals are prominent.

Most of these felsic rocks are altered. All have a microcrystalline goundmass containing phenocrysts of sodic to calcic oligoclase that are cloudy with inclusions of kaolin, sericite, and chlorite. Some contain rounded, embayed quartz phenocrysts. In some of these rocks irregular radial concentrations of dark indeterminate material are scattered throughout the goundmass.

Volcanic sedimentary and pyroclastic rocks comprise more than half the Silver Pass unit. Volcanic sedimentary rocks dominate the succession. They are drab brownish green, weather to brown, and are thick-bedded.

The volcanic sedimentary rocks are exposed east of Kachess Lake. They are composed mainly of volcanic fragments and might be called tuff and breccia; however, microscopic examination reveals much sedimentary material. The fragments in these rocks range from volcanic ash to blocks more than

10 feet in their greatest dimension. The finer-grained rocks consist of very poorly sorted, angular, altered volcanic chips mixed with quartz and plagioclase clasts and cemented with calcite. Volcanic sandstone occurs as both thick and thin beds and forms the matrix for the volcanic conglomerate. The volcanic conglomerate is also poorly sorted and composed predominantly of andesitic fragments averaging 6 inches in diameter. Some beds have angular fragments, others have rounded fragments, and a few have both. These beds, where well exposed, average about 20 feet in thickness.

At a few localities, especially in the vicinity of Silver Pass, there are a few thin beds of indurated arkose, composed of quartz, plagioclase, pyroxene, chlorite, and a few fragments of solidified lava.

In addition to the volcanic sedimentary rocks, some true pyroclastic rocks were noted, especially east of Silver Creek. These rocks are lithic lapilli-tuffs and are composed of fragments of altered andesite.

Teanaway basalt.—G. O. Smith (1903, p. 15) proposed the name Teanaway basalt for Russell's lowest or first sheet of Columbia lava. In his reconnaissance, Russell (1899, p. 129–131) used the term Columbia lava for all the Tertiary basalts.

In 1904, Smith (p. 5–6) stated

"... the name Teanaway has been applied to this formation, which includes only the basalt flows and interbedded basaltic pyroclastics of Eocene age and which constitutes a series that can be taken as a unit, since it represents the products of volcanic activity uninterrupted by any other important geologic process."

In 1906, Smith and Calkins (p. 5) noted "The Teanaway basalt has been fully described in the Mount Stuart folio. It was named for the river in whose basin it is most extensively developed." This is the only statement that could be considered to indicate a type area.

OCCURRENCE: The Teanaway basalt crops out in an arcuate band, up to 4 miles wide, extending from east of Kachess Lake to Table Mountain. It lies unconformably above the Swauk and concordantly below the Roslyn. The arcuate outcrop pattern is determined by relatively gentle folding that formed the Roslyn basin. The area discussed herein contains the main body of the Teanaway basalt; however, small exposures have been noted south of Cle Elum (Smith, 1904) and east of Table Mountain (Saunders, 1914, p. 152).

The Teanaway basalt ranges from about 1000 to more than 5000 feet thick and is composed of basaltic lava complexly interbedded with sedimentary and pyroclastic rocks. The general distribution of rock types within the Teanaway and their relationship to the dike swarm is shown by Foster (1958, Fig. 2, p. 646). This diagram shows how different this unit is from the Yakima basalt, in which individual flows can be traced tens of miles (Waters, 1955b, p. 708). The exposures of the Teanaway essentially represent a cross section through a volcanic field. The attitudes on Plate 1 show a marked discordance with the outcrop pattern and are probably influenced by initial dip. This is interpreted to mean that these rocks were originally deposited on the slopes of volcanoes.

Cross sections through the Teanaway basalt are afforded by the many streams that cross the trend of this unit. Smith intended that the type area be on the Teanaway River; therefore, the rocks along the Middle Fork of Teanaway River should be considered the type. The complex intermingling of lavas and clastic rocks, many with initial dips, make a section described and measured here of little value in characterizing the sequence.

In the type area along the Middle Fork of the Teanaway River, the Teanaway basalt is more than 5000 feet thick and is composed of basaltic lava, tuff, and lapilli-tuff. The upper part of the section is primarily clastic with tuff and lapilli-tuff predominating, although there is some columnar basalt.

In the narrow canyon of the West Fork of the Teanaway River are good exposures of interbedded lavas and clastic rocks. The ridges between here and Cle Elum Lake are composed almost entirely of breccia and lapilli-tuff.

The clastic rocks are also exposed along the county road on the east shore of Cle Elum Lake. At the north end of the exposures along the lake, a thin-bedded, east-dipping, micaceous sandstone is interbedded with green tuffs. Above these rocks are conglomerates with a sandy matrix and pebbles of both quartzite and basalt. This sequence is probably the base of the Teanaway; however, these rocks have apparently been jostled by the faults that trend nearly parallel to the lake, so their exact relationship is not clear.

The Teanaway rocks on the opposite sides of Cle Elum Lake differ in rock type and are offset. These features are interpreted as further indications of faulting. On the west side of the lake, the Teanaway is tightly folded, aphanitic basalt.

To the east, a good section is exposed in Swauk Creek and the roadcuts along U. S. Highway 97, south of Liberty. Here the lower part of the Teanaway is mostly a dark aphanitic to glassy basalt that has an ellipsoidal aspect in

some places. Some of the lavas are associated with clastic rocks. The upper part of the section here is predominately clastic and contains green or yellow-green tuff and lapilli-tuff. East of Swauk Creek and U. S. 97, the clastic rocks are exposed on dip slopes north of First Creek. On the west face of Table Mountain, below the Yakima basalt, interbedded clastic rocks and lavas are exposed.

For petrographic descriptions of the Teanaway rocks, *see* Foster (1958, p. 648–649).

RELATIONSHIPS: The Teanaway basalt lies unconformably above the previously folded Swauk formation and, locally, above the Silver Pass volcanic rocks. The Teanaway is overlain by the Roslyn formation, the Yakima basalt, and what is here called post-Roslyn rhyolite. The Roslyn, oldest of the three formations, concordantly overlies the Teanaway; the other two are younger and are unconformable on the Teanaway.

Early workers agreed that the Roslyn concordantly overlies the Teanaway; however, the exact nature of the contact is controversial. Russell (1899, p. 123) described it as gradational so that "... no two observers would draw the boundary in precisely the same place." In a footnote (1899, p. 123) he notes that Willis "... states that the base of the Roslyn on the Teanaway is precisely determined by a conglomerate containing pebbles of basalt." Smith and Calkins (1906, p. 6) stated "It [the Roslyn] overlies the basalt without observed angular unconformity or distinct evidence of an erosional interval."

C. T. Bressler (1951, Unpub. PhD thesis, Pennsylvania State College, p. 28) noted a red bed about 200 feet thick at the contact between the Teanaway and the Roslyn. He interpreted this red bed and a discordance he observed at the mouth of the canyon of the Middle Fork of the Teanaway River to mean that the Roslyn is disconformable on the Teanaway.

AGE AND CORRELATION: A few fragments of *Viviparus* were found in beds exposed at very low water level on the east shore of Cle Elum Lake, just south of Morgan Creek. The rocks here are thin-bedded sandstone and shale, with a few conglomerate beds composed of rounded mudstone fragments up to 10 inches in diameter. All the *Viviparus* fragments were found in a single fragment from this conglomerate. These mollusks were studied by Dr. J. A. Aurèle La Rocque, who was unable to speciate them due to their poor preservation. The genus *Viviparus* ranges from at least Cretaceous to Recent. At the same general locality, possible worm borings were also noted. The only other organic materials noted in these rocks were a few macerated leaves and stem material.

The Teanaway basalt unconformably overlies the Swauk formation, of probable Late Cretaceous or Paleocene age and is overlain concordantly by the Roslyn arkose of probable middle or late Eocene age (Wheeler, 1955, p. 1668). These stratigraphic relationships suggest a tentative Eocene date for the Teanaway basalt.

Dike swarm and diabase associated with Teanaway basalt.—The Teanaway dike swarm is one of the most impressive features in this area (Fig. 2). The following description is summarized from Foster (1958).

The swarm extends from the area of this report 45 miles east to Wenatchee. Its best development is in the area south of the Mount Stuart block. Here, at places, dike material forms up to 90 per cent of the surface area, although the average is much less. The dikes are nearly vertical and range from 15 to more than 150 feet in thickness, and a few can be traced for several miles. Most of the dikes intrude the Swauk arkose, but some intrude all the pre-Teanaway rocks such as the Silver Pass volcanic rocks, Easton schist, and the peridotite. The dikes are perpendicular, or nearly so, to the fold axes, and the swarm is best developed where the folding is the tightest. This suggests that the forces that folded the Swauk and older rocks controlled the emplacement of the dike swarm, the dikes being emplaced parallel to the direction of least compression.

A related diabase body near Dry Creek is shown on Plate 1. Several other bodies, too small to show on the map, were noted. The diabase appears to be a coarser-grained equivalent of the dike rock.

Earlier workers believed that the dike swarm provided the vents through which the Teanaway basalt was extruded. This is not borne out by detailed mapping. The attitudes within the Teanaway basalt are discordant with the outcrop pattern and suggest extrusion from a few widely spaced volcanic centers rather than the closely spaced dike swarm. Petrographically the dike rocks are indistinguishable from the Teanaway basalt lavas (Foster, 1958, p. 646–649).

Roslyn arkose.—The Roslyn formation was named by Russell (1899), and no type section has ever been designated. The formation was studied in detail by C. T. Bressler (1951, unpub. PhD thesis, Pennsylvania State College).

The Roslyn is deformed in the area just south of the map area of this paper into an asymmetrical syncline with the steep limb on the southwest side. The west side of this syncline is tightly folded and faulted. The Roslyn is poorly exposed on low wooded hills and ridges and better exposed on the forks of the Teanaway River.

C. T. Bressler (1951, Unpub. PhD thesis, Pennsylvania State College, p. 29) divided the Roslyn into three units. He noted that the basal beds of the Roslyn consist of 200 feet or less of red clastic rocks that grade upward into nearly white clastic rocks. Above the basal beds is 2300 feet of largely massive, yellowish-gray, arkosic, medium-grained sandstone with darker, interbedded, fine-grained sandstone and siltstone. The upper unit is 1500 feet thick and is composed of eight major coal beds interbedded with medium- to fine-grained sandstone and siltstone. Saunders (1914, p. 28) reported interbedded lava sheets and tuff beds in the Roslyn. He gave no locations, and no other workers have reported these volcanic rocks.

The Roslyn in the area studied is in contact with two younger rocks, the post-Roslyn rhyolite and the Yakima basalt. The relationships with the rhyolite are considered hereafter. The Yakima basalt overlies the Roslyn with an angular unconformity that can be seen on the west face of Table Mountain.

The age of the Roslyn has been repeatedly referred to because it is the only reasonably accurately dated formation in the column. It is considered middle and/or upper Eocene on the basis of one fossil fish and two turtle carapaces, which, together with leaves, are the only fossils found in these rocks to date (Wheeler, 1955, p. 1668). The original leaf collection was studied by F. H. Knowlton, who reported (*in* G. O. Smith, 1904, p. 7) an Eocene age younger than the Swauk, noting that not a single species was common to the Swauk. The fish was studied by

FIGURE 2.—TEANAWAY DIKES

A shows dikes on the southeast flank of Jolly Mountain. The light areas are the septa of Swauk sediments. The figure in the foreground gives the scale. *B* shows almost flat-lying Swauk sediments intruded by Teanaway dikes on the ridge north of Sasse Mountain.

Hesse (1936), who considered it probably middle Eocene.

Post-Roslyn rhyolite.—The post-Roslyn rhyolite is a group of altered felsic volcanic rocks. They are scattered in an area a few miles wide bordering the outcrop of the Teanaway basalt and are poorly exposed in this wooded country. Russell (1899, p. 130) wrote,

> "In intimate association with these [the Teanaway basalt] in the region embraced between the North and Middle Forks of the Teanaway River, there are peaks composed of nearly white rhyolite. The relation of the rhyolite to the Columbia basalt [Teanaway basalt] is not well known, but it seems to be a dike exposed by erosion and much shattered and altered in color by weathering."

G. O. Smith wrote (1903, p. 18)

> "... lavas occur on the middle fork of the Teanaway River, where rhyolitic rock is seen to rest upon the eroded surface of the older rocks. The age of this lava is somewhat problematical, but evidently it is considerably younger than the Eocene rocks with which it is in contact."

Smith (1904, p. 8) gave a more complete description of the rocks and their relationships. He described the large rhyolite occurrence west of Ryepatch as a post-Teanaway flow. Elsewhere, however, he noted that the rhyolite overlies both the Roslyn and the Teanaway, and suggested a Pliocene age. No name has ever been suggested for these rocks and, because their relationships are obscure, it seems best to refer to them as post-Roslyn rhyolite.

Each of the scattered occurrences must be studied as a separate unit, for rocks of more than one age may be present.

These rocks are all altered, which commonly makes them weather easily and so obscures their relationships to the surrounding rocks. It is commonly difficult to obtain a solid specimen, because the altered rock crumbles under the hammer. Smith (1904, p. 8) observed that some of the rhyolite resembled altered shale in the field.

DESCRIPTION: These rocks range from creamy white to very light brown; many have an earthy appearance and are limonite-stained. Some of the rocks have clastic textures shown by color differences between the fragments and the groundmass. In a few rocks, small spherulitic growths can be seen with a hand lens.

Two tiny occurrences on First Creek are welded tuff. The groundmass in these rocks is altered and somewhat devitrified. The outlines of flattened glass shards are emphasized by concentrations of limonitic material. In addition, these rocks contain a few phenocrysts of quartz and altered feldspar.

At other localities, the rocks are so altered that little can be said about their origin. In some, spherulitic growths obscure the earlier textures. Other rocks consist of a plexus of amygdules of silica or calcite. In others, the alteration to clay minerals, sericite, limonite, and alunite has proceeded to a point where the original textures cannot be determined.

Yakima basalt.—The Yakima basalt forms the east and southeast border of the area of this study.

The Yakima basalt is one of G. O. Smith's subdivisions (1901, p. 15) of Russell's Columbia lava. Smith defined the Yakima as the basalt of the Miocene period [*sic*] and designated no type section. Yakima basalt is used in this study as Smith used the term in practice, that is, for the pre-Ellensburg formation basalts.

The Yakima basalt is composed of a large number of flows, some of which can be traced for many miles. The typical rock is an aphanitic to fine-grained black or dark-gray basalt with a few phenocrysts of feldspar.

The age of this formation is near the Miocene-Pliocene border (Waters, 1955a, p. 664). It is reasonably well dated by leaves, fresh-water mollusks, and vertebrates found in the sedimentary rocks between the Yakima basalt flows and especially in the Ellensburg formation.

The Yakima basalt is relatively resistant, and where it overlies the Swauk and Roslyn formations, large landslides have developed as a result of erosion of the weaker underlying rocks. The resulting landslide topography along the face of Table Mountain and Lookout Mountain is striking.

Howson andesite.—The Howson andesite occurs only in two small outcrops on Sasse Mountain north and northeast of Howson Creek. It was named and described by G. O. Smith and Calkins (1906, p. 10), and much of the following was abstracted from their description. It is a light-gray andesite with prominent elongate hornblende phenocrysts and smaller feldspar phenocrysts. It overlies Swauk formation that has been intruded by Teanaway dikes. The Howson andesite has well-developed columnar structure which gives rise to extensive talus slopes that obscure the contact with the older rocks.

The Howson andesite, unlike the other andesites in the area, is unaltered. It is very similar to the hornblende andesite boulders in the Ellensburg formation, and Smith and Calkins suggested that it may be the source of the Ellensburg detritus. This may put an upper limit on the age of the Howson, for the Ellens-

burg detritus seems to have come from active volcanoes. The age of the Ellensburg formation lies near the Miocene-Pliocene border (Waters 1955a, p. 664).

Smith and Calkins report zoned plagioclase, hypersthene (in some specimens), and hornblende of several varieties with differing optical properties, in a groundmass of glass, iron oxide, quartz, and orthoclase.

The Howson andesite is apparently the youngest formation in this and the surrounding area.

Western Stratigraphic Section

Denny formation.—These rocks were originally called Guye formation by Smith and Calkins; however, detailed mapping has shown that they are older than the Guye formation and probably pre-Tertiary.

OCCURRENCE AND RELATIONSHIPS: The Denny formation is confined to two small areas on Denny and Snoqualmie Mountains with a total extent of less than 2 square miles. Coarse-grained marble is the most distinctive rock in the Denny formation, and it is interbedded with hornfels. W. R. Danner (1957, unpub. PhD thesis, Univ. of Washington, p. 249) reports associated basalts, perhaps of Tertiary age, on Denny Mountain. There are no exposures between the Naches hornfels in the Snoqualmie River and the Denny rocks on Denny Mountain, and a fault seems likely.

The Denny formation is well exposed on the south flank of Denny Mountain and in the Guye Peak and Snoqualmie Mountain saddle. The last location, because of its better exposures, is here designated the type area. At this locality, the rocks are medium to coarse crystalline marbles interbedded with fine-grained hornfels. The intrusion of the Snoqualmie granodiorite is responsible, at least in part, for the metamorphism. The thickness cannot be estimated because the deformation and metamorphism that accompanied the emplacement of the Snoqualmie granodiorite have obscured the structure. In this relatively flat saddle, karst topography has developed, and many sinkholes were observed.

Bedded chert on Chair Mountain is shown on Plate 1 as Denny formation.

AGE AND CORRELATION: The rocks of the Denny formation were originally considered as Miocene with some reservation by Smith and Calkins, who included them in their Guye formation. The field relationships show that they are pre-Snoqualmie granodiorite and probably in fault contact with the other rocks. These rocks are believed to be pre-Tertiary because no limestone has been found in the Tertiary rocks of the Cascades except for a few, thin, limited beds of fresh-water limestone in the Puget formation. W. R. Danner (1957, unpub. PhD thesis, Univ. of Washington, p. 249) believes that these rocks are lithologically similar to the Permian rocks of the Northern Cascades.

These rocks alternatively may be the southeastern end of a group of metamorphic rocks reported by H. L. Bethel (1951, unpub. PhD thesis, Univ. of Washington, p. 78–81) or a graywacke unit Bethel described in which Danner (1955, p. 454) found Jurassic-Cretaceous fossils.

DESCRIPTION: The metamorphism of the Denny formation appears to be due to the emplacement of the Snoqualmie granodiorite because (1) the metamorphism is static, with hornfelsic textures predominant, and (2) the grade of metamorphism increases as the granodiorite is approached. The Denny rocks do not occur at any great distance from the granodiorite, so the contact metamorphism obscures any earlier metamorphism.

In hand specimen, the marbles are medium- to coarse-grained, are mottled and are light gray and weather to light or medium gray. In thin section, these marbles show a mosaic of anhedral calcite crystals. In some thin sections, clusters of small diopside crystalloblasts were observed.

The composition of the hornfels is variable. In hand specimen most of the hornfelses are medium- to dark-gray or greenish-gray aphanitic rocks that weather to medium or light gray. The hornfelses at some distance from the granodiorite are very fine-grained, and the minerals are commonly turbid. These rocks are composed of quartz, a little oligoclase, epidote, opaque minerals, and in some specimens, actinolite, chlorite, or hornblende. The higher-grade rocks are more coarsely crystalline and are characterized by diopside and calcic andesine, commonly with sphene and apatite as abundant accessories.

The contact between the marble and the Snoqualmie granodiorite is the site of local skarn mineralization. This mineralization consists of small pods of magnetite in the marble and massive garnet bands, garnet crystals, pyrite, specular hematite, actinolite, quartz crystals, and epidote (Shedd, Jenkins, and Cooper, 1922).

Guye formation (restricted).—The name Guye formation was applied by Smith and Calkins (1906, p. 7) to a heterogeneous group of sedi-

mentary and volcanic rocks whose structure was not clearly known. As a result, rocks of several ages were included in the formation. This composite unit was erroneously dated Miocene from a small collection of fossil leaves, although Smith and Calkins, and most of the later workers, realized that the rock types and the structure resembled the lower Tertiary rocks more closely than the Miocene rocks. The Guye formation was one of the few dated units in the region, so its age was used to date more widespread formations. The age of the Guye formation became an enigma, and many papers on Cascade Tertiary rocks discussed it, but no one studied the rocks themselves.

The Guye formation of Smith and Calkins has been separated here into four formations—the Denny formation, the Guye formation (restricted), the Mt. Catherine rhyolite, and the Naches formation. Hereafter, unless otherwise noted, all reference to the Guye formation will be to the restricted Guye formation, which is composed of shale, sandstone, and conglomerate.

OCCURRENCE AND RELATIONSHIPS: The Guye formation crops out in only two areas, one northwest and the other south of Snoqualmie Pass. The larger of these areas is the ridge that extends south from Snoqualmie Pass. This exposure is bounded on three sides by the Mt. Catherine rhyolite, which dips away from the Guye on all sides and apparently unconformably overlies the Guye. From these relationships the structure is interpreted as an anticline in the rhyolite with homoclinal Guye formation exposed in the core. Faulting could also produce this outcrop pattern, but because of the very poor exposures, little more can be said. Throughout this area, the Guye strikes approximately N. 30° E. and dips 45–70° southeast. Rare cross-bedding shows the top is to the southeast. The exposures are generally poor in this area of heavy vegetation, but outcrops can be found in creeks, scattered along the ridge crest, and in road cuts. As there are no large areas of continuous outcrop, the most accessible place to designate a type area for the restricted Guye formation is on Coal Creek and its tributaries, on the east side of Snoqualmie Pass. These exposures display the distinctive Guye rocks, black leaf-bearing shales, in part carbonaceous, interbedded with dark coarse- and fine-grained sandstone which abruptly interfingers with conglomerates of subangular fragments of black and gray chert. This type area almost parallels the strike of the Guye, so that its main disadvantage is the limited stratigraphic exposure.

The second area of Guye outcrop is on Denny and Snoqualmie mountains, northwest of Snoqualmie Pass. On the high ridges in this area, indurated Guye sedimentary rocks are exposed, dipping southeastward at about 50°. On Denny Mountain, the Guye is overlain, apparently unconformably, by the Keechelus andesite breccia that forms the summit ridge. Elsewhere in this area the Guye is intruded by the Snoqualmie granodiorite, except locally on Chair Mountain where the Guye overlies chert beds (W. R. Danner, 1957, unpub. PhD thesis, Univ. of Washington, p. 250).

AGE AND CORRELATION: Leaves are the only fossils found to date in the Guye formation. The first report on these leaves was by F. H. Knowlton (*in* Smith and Calkins, 1906, p. 7), who reported

> "*Platanus dissecta* Lesq.
> *Acer aequidentatum* Lesq.
> *Ficus* n. sp., cf. *F. artocarpoides* Lesq.
> *Cinnamomum* n. sp.
>
> "The first two species are without question Miocene. The *Ficus* approaches quite clearly *F. artocarpoides*, which is from the Fort Union group (Eocene) [*sic*] of Montana. The specimen is not perfect and cannot therefore be positively determined. If it differs at all from the Fort Union species the difference must be very slight. Yet the species may prove to be new.
>
> "The *Cinnamomum* is undoubtedly new and is a fine characteristic species. It does not approach any described species very closely, yet appears to be of a Tertiary type.
>
> "The two species that can be determined are Miocene, and the other two are of Tertiary, probably Miocene facies."

Smith and Calkins commented on the determination, saying,

> ". . . the stratigraphic relations to the overlying rocks, added to the lithologic resemblance of the Guye formation . . . to the Eocene formations, would have led to its reference to the Eocene were it not for the paleobotanic evidence just cited."

Beck (1934, p. 3) correlated the Guye with the Naches formation and suggested that both the Naches and Swauk may be Cretaceous.

R. W. Brown, in a restudy of the original Knowlton collection, (*in* Warren, 1941, p. 810–811) noted that the two Miocene forms were probably ". . . leaf variations of a single species, probably ancestral to *Platanus dissecta* Lx", and that the *Ficus* was ". . . too fragmentary to identify satisfactorily". He considered ". . . the excellent specimens of *Cinnamomum* are very similar to, if not identical with, *C. dilleri* Kn."

Dr. Brown very kindly studied a leaf collection made by the author from the south-

western bank of Coal Creek, a few hundred feet below the mouth of Hyak Creek. He reported (Personal communication, June 8, 1954):

> "*Allantodiopsis erosa* (Lesquereux) Knowlton
> *Asplenium magnum* Knowlton
> *Glyptostrobus dakotensis* Brown
> *Ocotea eocernua* Chaney and Sandborn
> "Comment: This is an Eocene assemblage, and probably from the later half of the Eocene. Most of the specimens in the collection represent the one species of *Ocotea*."

Brown's determinations are summarized in the following chart. The ranges are from LaMotte (1952).

	Paleocene	Eocene	Oligocene
Allantodiopsis erosa (Lesquereux) Knowlton	×	×	..
Asplenium magnum Knowlton	×	..	..
Cinnamomum dilleri Knowlton	..	×	..
Ficus artocarpoides Lesquereux	×	×	..
Glyptostrobus dakotensis Brown	×	×	..
Ocotea eocernua Chaney and Sanborn	..	×	×

On this basis the Guye formation may be Paleocene or Eocene. Some of these leaves are found in the marine Eocene rocks in southwestern Washington (Snavely *et al.*, 1951; Roberts, 1958). Bell (1956, p. 110) reports *A. erosa* in Upper Cretaceous (Senonian) rocks in British Columbia. He suggests that some Tertiary plants appeared first on the Pacific coast and then migrated inland.

The Guye is unconformably overlain by the Mt. Catherine rhyolite, which is overlain, apparently concordantly, by the Naches formation of probable Eocene age. Thus the Guye formation is unconformably below rocks that may be Eocene, suggesting an Eocene or Paleocene age for the Guye.

DESCRIPTION: The Guye formation is a thick sequence of shale, sandstone, and conglomerate. Its fluviatile origin is shown by the many lenses of conglomerate that are interbedded with the sandstone and leaf-bearing shale. The thickness appears to be at least 5000 feet, but sparseness of outcrops prevented detection of any repetition by faulting.

The conglomerate, sandstone, and shale are mixed in all proportions, and the beds or lenses range from thin to massive. There are no key beds.

Conglomerate, percentage-wise, is the least important rock type of the Guye formation, but it is the most distinctive. It is composed of angular to subangular fragments, up to 2–3 inches in diameter, of medium-gray to black chert and white quartz. These fragments are set in a matrix of coarse- to fine-grained sandstone, similar to that described hereafter.

The chert fragments have tiny quartz veins, and the quartz fragments appear to be vein quartz.

The sandstone ranges from medium gray to dark greenish gray, depending, in part, on the amount of carbon, and weathers gray or brown. The grain size ranges from fine to very coarse. Fragments of quartz and feldspar can be identified with the hand lens, and carbonaceous layers or fragments mark the bedding in some outcrops. These rocks are well cemented, but near the granodiorite they become silicified and very indurated.

These sandstones contain about 50 per cent quartz or chert, 20–40 per cent feldspar, up to 30 per cent chlorite, up to 10 per cent each of biotite and opaque materials, and up to 5 per cent muscovite. Three-quarters of the feldspar is plagioclase, much of which is twinned. Most of the feldspar is turbid, but some is clear. Most of the chlorite has replaced biotite, which has partially altered to chlorite and magnetite. The opaque materials appear to be iron oxides and carbon. Scattered grains of other minerals are present, and Wilson, Skinner, and Couch (1942, p. 57) report augite, rutile, hornblende, garnet, and glauconite. Glauconite is usually interpreted as indicating a marine origin; however, the present author found none in his thin sections.

The grains in these sandstones are angular to subangular. They are poorly sorted, and most have a chloritic matrix. Some of the grains in a few of these sandstones have sutured boundaries.

The Guye shales and siltstones are black or very dark gray and weather light or medium gray. Some of these shales contain pyrite. Fossil leaves are crowded on bedding planes at many horizons. In many areas, these shales are coaly, but no coal beds of any extent have been found. Many of these shales contain very fine grained sand, and sandy partings on bedding planes are not uncommon. At a few localities, pebbles or granules of quartz or chert in the shale are common.

The Guye rocks near the Snoqualmie granodiorite are indurated and silicified. Even near the granodiorite, the alteration of the quartz-rich sedimentary rocks, especially the chert-pebble conglomerates, consists only of silicification and some recrystallization. The fine-grained cement or matrix of the sandstones and conglomerates is most affected and becomes quartz, chlorite, sericite, and opaque materials, presumably graphite and iron oxides (Goodspeed, 1947, p. 65). Tourmaline porphyroblasts are present at places near the granodiorite, and garnet porphyroblasts were noted in the matrix in several thin sections. The origin of the garnet is probably similar to that of the garnet-quartz-diopside veinlets described by Goodspeed and Coombs (1932, p. 554). A few of these sandstones are completely altered to hornfels in which the original clastic texture is replaced by a granoblastic texture.

Mt. Catherine rhyolite.—The Mt. Catherine rhyolite, a new formation, is composed of light-purple, light-orange, and light-blue-gray rhyolite flows, tuff, and breccia, most of which contain small clear quartz phenocrysts. Smith and Calkins (1906) showed only the southern part of the outcrop of this rhyolite on their map. The outcrop pattern of the Mt. Catherine rhyolite is a crude oval, open to the northwest.

It is not well exposed; the best exposures are on the southeast side of the oval from Gold Creek to Mt. Catherine. A nearly complete section can be seen in the steep-walled canyon of Mill Creek, the southeasterly running creek just northeast of Mt. Catherine. Mill Creek, although access to it is difficult because of the narrowness of the canyon, should be considered the type area.

The Mt. Catherine rhyolite overlies the Guye formation with an angular unconformity. The rhyolite dips away from the inside of the arc at moderate to steep angles, and the Guye sedimentary rocks dip steeply southeast, so the rhyolite is parallel with the Guye on the southeast side of the arc but overlies it discordantly elsewhere. The Mt. Catherine rhyolite is overlain, apparently concordantly, by the Naches formation. This relationship can be seen on Kendall Peak and especially in Cold Creek.

AGE AND CORRELATION: The Mt. Catherine rhyolite contains no fossils; however, it unconformably overlies the Guye formation of probable Paleocene or Eocene age and is concordantly overlain by the Naches formation of probable Eocene age. Thus it appears that the Mt. Catherine is probably Eocene.

DESCRIPTION: The Mt. Catherine rocks are, in general, easily recognized. Most hand specimens are some shade of purple, although some are light orange or light blue gray, and most have small clear quartz phenocrysts. The weathered rock is buff and commonly limonite-stained. Some of the rocks are flow-banded, and others are clastic.

Clastic rocks appear to be most common, at least in the places where alteration has not obscured the original texture. There are two types of clastic rocks. The first is exposed on U. S. 10, just north of Gold Creek. In hand specimen, this rock is light purple and has many small clear quartz phenocrysts. It is composed of angular fragments of purple porphyry set in a matrix of the same rock. Both the fragments and the matrix are composed of devitrified glass. Embayed quartz occurs in both the fragments and the matrix. The fragments are angular and range from about ½ mm to more than 1 cm. There are a few phenocrysts of altered plagioclase, probably albite, orthoclase altered to sericite, and ferromagnesian minerals altered to chlorite.

The other common clastic rock in the Mt. Catherine rhyolite is bedded tuff that commonly occurs near the base. It is composed of angular fragments in a matrix of what appears to be streaked-out glass. In thin section, this rock is seen to be composed of devitrified glass with a few possible glass shards and a few rock fragments, mostly of a trachytic lava. The devitrification masks the texture, so little can be said about the origin of the rock. This basal tuff is sheared in the area where Coal Creek crosses the rhyolite and was exposed in the foundation excavations for the bridge built in 1956.

The flow-banded rhyolites have thin contorted bands and are otherwise similar to the rocks described heretofore. They are devitrified and contain embayed quartz phenocrysts and a few feldspar crystals altered to sericite. At a few localities, these lavas contain calcite amygdules.

At many localities, especially to the south and west, the original texture of the Mt. Catherine rhyolite is obscured by later spherulites. These spherulites show a dark cross under crossed nicols, and no structure can be seen in plane light except concentric zones of tiny inclusions. The fine-grained matrix between the spherulites is apparently composed of quartz and chlorite. Perlitic cracks occur with the spherulites in some rocks.

Naches formation.—The Naches formation was named by Smith and Calkins (1906, p. 4–5) who stated

> "The Naches formation is composed of interbedded sandstone and basalt, the sedimentary rock predominating in the lower, and the volcanic in the upper portion. The formation is named for the river in whose basin it is most extensively developed."

Smith and Calkins believed that the Swauk and the Naches were of the same age on the basis of two fossil leaves found in both and that they were laid down in separate contemporaneous lakes. However, abundant evidence indicates that these rocks are fluviatile, and, in spite of the fossil leaves, the complete lack of contemporaneous volcanic rocks in the Swauk suggests that they are of different ages. These two formations are not in stratigraphic contact. Smith and Calkins did show that both Swauk and Naches unconformably overlie Easton schist. They also showed that in the southern part of Snoqualmie quadrangle, the Keechelus volcanic rocks overlie the Naches formation unconformably.

The rocks that are correlated with the Naches formation herein were called Teanaway basalt, Guye formation, and Keechelus formation by Smith and Calkins.

OCCURRENCE AND RELATIONSHIPS: In the western part of the area of this study, there is a thick, widespread unit of interbedded sedimentary rocks and basalt with a few rhyolitic

rocks that is similar to the Naches formation of Smith and Calkins. This unit is apparently overlain unconformably by the Keechelus andesite and locally near Snoqualmie Pass is underlain, concordantly, by the Mt. Catherine rhyolite. These rocks are correlated with the Naches formation of Smith and Calkins because of lithic similarity, apparently similar structural and stratigraphic position, and because they are both thick widespread units almost in contact on opposite sides of the Yakima Valley.

Smith and Calkins called the rocks west of the Kachess lakes Teanaway basalt, probably because basalt is the most common natural outcrop in this area, but they did observe the intercalated sedimentary rocks. Current logging operations have exposed nonresistant sedimentary rocks. Shale, coarse- and fine-grained commonly arkosic sandstone, and conglomerate comprise about half the Naches. The Teanaway basalt formation contains very few sedimentary rocks.

The structure in the area between Lake Keechelus and Little Kachess Lake is crucial for determination of the relationship between the Naches and the Keechelus. West of Little Kachess Lake the Naches rocks strike northwesterly and dip steeply. The Naches fold axes shown on Plate 1 are diagrammatic because it was not possible to determine the stratigraphy of the formation. East of Gold Creek the Keechelus rocks strike northerly and northeasterly and dip eastward at moderate angles. It is difficult to find the contact between these two units in the field because in the vicinity of the contact the Naches and Keechelus lavas are similar and there are few exposures of Naches sedimentary rocks. Farther south in the area north of Keechelus Lake it is impossible to trace the contact because of the forest cover. Also, there may be a difference in the nature of the Keechelus. It is interesting to compare the present map with that of Smith and Calkins (1906) because both maps show a moderately dipping contact that becomes steep in the area north of Keechelus Lake. As noted in the discussion of the Silver Pass volcanic rocks, M. L. Stout (1959, Unpub. PhD thesis, Univ. of Washington), working south of the present area, found andesites in the upper part of the Naches formation. If these andesites are present in the area west of Little Kachess Lake, they may account for the difficulties just described. Study of the structure on Plate 1 shows that the fold axes may cross the Naches-Keechelus unconformity. In spite of the problem of accurately mapping the Naches-Keechelus contact, there are distinctions, both structural and lithologic. In the area north of the present study, R. C. Ellis (1959, Unpub. PhD thesis, Univ. of Washington, p. 81–83) has followed this contact to the crest of the Cascade Range.

To the east, at Little Kachess Lake, the fold axes terminate against Easton schist. The contact is interpreted as a fault, the Kachess fault. This fault has been mapped nearly 10 miles north of Little Kachess Lake by R. C. Ellis (1959, unpub. PhD thesis, Univ. of Washington), who reports that on the Cascade crest between Chimney Peak and Summit Chief Mountain, the fault plane is exposed. Here the fault dips about 65° east.

The rocks on Amabilis Mountain are considered Naches formation, although the poor exposures prevent close correlation and detailed mapping. The only outcrops are along the lake shore, scattered on the hillsides, and in the recent highway cuts. Exposed on the lake shore are arkose to the south and rhyolite to the north. The road cuts expose sedimentary rocks and a variety of igneous rocks, including basalt dikes, possibly part of the Teanaway swarm.

To the north, the Naches formation is intruded by the Snoqualmie granodiorite. In the Mineral Creek region, this contact is the site of copper mineralization.

The Naches in the vicinity of Snoqualmie Pass has, for the most part, been metamorphosed to hornfels by the Snoqualmie granodiorite. The rocks exposed along U. S. 10 on the west side of Snoqualmie Pass are mainly basalt hornfels, in a few of which the original textures are still visible. Smith and Calkins were confused by the rocks on Kendall Peak and described several puzzling rocks, as for example (1906, p. 7),

> "A very peculiar apparently rhyolitic rock is found on the southwest shoulder of Kendall Peak. It consists mainly of a rather hard, aphanitic groundmass, dull coal black in color, which contains abundant small angular grains and crystals of quartz. Its texture and composition studied microscopically, indicate that it is tuffaceous. Its black color is due partly to an abundance of fine scaly material resembling green biotite and numerous black opaque particles of undetermined character.
>
> "The basalt near Kendall Peak is mostly a greenish black, compact aphanitic rock, not very readily distinguished from the indurated black shale. Its true character was first recognized by the finding of amygdaloidal phases, with cavities full of quartz, hornblende and other secondary minerals. A little indurated tuff was also found."

AGE AND CORRELATION: The only described fossils in the Naches formation are leaves in

carbonaceous shale on Gale Creek, although wood fragments are ubiquitous. Poorly preserved leaf fragments were observed at several localities, but none were suitable for identification. A collection from the carbonaceous shale near the mouth of Cold Creek was examined by Dr. R. W. Brown, who reported only undiagnostic leaf fragments and water-lily root scars. The Gale Creek flora was examined by F. H. Knowlton who reported (*in* Smith and Calkins, 1906, p. 6):

> "This is a collection consisting of a dozen or more pieces of matrix showing numerous impressions. I note the following genera: *Populus* sp., *Ficus* sp., *Pterospermites*, sp., *Ulmus* (?) sp.
>
> "I do not recognize any of the species in this collection. The genera are sufficiently clear, but the species are new and are unlike any heretofore obtained. The forms are not similar to those found in either the Roslyn or the Swauk, but appear to find their greatest affinity with material from the west side of the Cascades, at Carbonado and vicinity. For example, no Eocene species of *Populus* has heretofore been found east of the Cascades, while several forms have been found about Carbonado, South Prairie Creek and Franklin. The species of *Ulmus* is similar to elm leaves from vein 12 at Franklin, but I am not certain that it is absolutely identical."

The suggested correlation of these rocks on the east side of the Cascades with the Puget formation at Carbonado and vicinity on the west side of the Cascades is somewhat supported by the relationships of the rocks on both sides of the Cascades. In the area of this study, the Naches formation is unconformably covered by the Keechelus andesite. To the west, only Keechelus rocks are exposed until the western front of the range is reached, where the Keechelus rocks give way to the underlying Puget formation. Lithologically, the Puget is similar to the Naches but contains more carbonaceous sediments and fewer volcanic rocks. The contact of Puget and Keechelus has been reported as almost concordant by Warren *et al.* (1945) and as interbedded by Fisher (1954, p. 1340), on the basis of doubtful correlations. A. T. Abbott (1953, unpub. PhD thesis, Univ. of Washington) reported Puget rocks in the Mt. Aix quadrangle south of the Snoqualmie quadrangle, but the evidence for this correlation is weak. The Puget is considered to be middle or upper Eocene.

Knowlton (*in* Smith and Calkins, 1906, p. 5) thought that the Naches correlated with the Swauk and that some of the rocks here called Naches formation (along Gale Creek) correlated with the Puget (Smith and Calkins, 1906, p. 6). Beck (1934, p. 3) reported the same leaf in beds here considered Naches and Swauk.

The Naches formation is underlain with apparent conformity by the Mt. Catherine rhyolite that unconformably overlies the Guye formation of possible Paleocene or Eocene age. The Naches formation is unconformably overlain by the Keechelus andesite that yielded an Oligocene or Miocene mammal fossil in the Mt. Aix quadrangle to the south. Thus the Naches may be Eocene. This dating and correlation should be considered tentative, however, until diagnostic fossils are discovered.

DESCRIPTION: The Naches formation is composed of approximately 5000 feet of interbedded basalt, sedimentary rock, and rhyolite. The stratigraphy varies from place to place, so that it is not possible to describe a general section, although basalt is usually more prevalent in the lower part.

Several types of basalt can be distinguished. Most of the basalt is amygdaloidal or porphyritic or both. In hand specimen, most types are green gray or gray and have prominent white plagioclase phenocrysts and/or white or green amygdules. These basalts commonly weather brown or red brown.

Porphyritic, amygdaloidal basalt is the most common. This group of basalts can be subdivided on the basis of texture. Those with an intergranular texture are green gray and have conspicuous dark-green and/or white amygdules. They average 55 per cent labradorite, 20 per cent augite, 20 per cent chlorite, and 5 per cent opaque minerals. The plagioclase has a composition of about An_{60} and is in the form of microlites and lath-shaped phenocrysts, many of which are fractured. The augite is faint violet and occurs as tiny grains. The opaque materials are for the most part interstitial, and the chlorite is in part interstitial and in part fills amygdules. Many of these rocks are altered, and much of the pyroxene is altered to chlorite and opaque minerals. In a few of these altered basalts the plagioclase phenocrysts are glomeroporphyritic, and the individual phenocrysts are commonly divergent. In the altered rocks, there are many amygdules filled with epidote, calcite, chlorite, and quartz. A few of these rocks contain the mineraloid iddingsite.

Some of the porphyritic, amygdaloidal basalts have an ophitic or a subophitic texture. In some of these rocks the plagioclase is labradorite, and in others it is as sodic as oligoclase.

The porphyritic, nonamygdaloidal basalts are similar except that they are less altered. Some are subophitic and others are pilotaxitic.

The nonporphyritic basalt is similar but less common. In general, those with amygdules are more altered. Some of these nonporphyritic rocks have an ophitic texture and are similar to the basalts described. They grade into trachytic and pilotaxitic

textures. Intermediate between the ophitic and the trachytic textures are subophitic rocks.

The Naches basalts exposed on the Sunset Highway west of Snoqualmie Pass have been thermally metamorphosed by the Snoqualmie granodiorite. These rocks have porphyritic or glomeroporphyritic plagioclase laths in a black to dark-gray, aphanitic groundmass. The plagioclase laths average three-quarters by three-sixteenths of an inch. Quartz-filled vesicles, about one-quarter inch in diameter, are present at some localities. These rocks are composed of fractured labradorite laths altered in part to sericite, chlorite, epidote, and opaque minerals, in a groundmass of smaller labradorite laths, chlorite, and actinolite. The actinolite is in the form of radiating crystals and veinlets. In some sections, epidote, calcite, stilpnomelane, clinozoisite, zoisite, apatite, and garnet are also present. These rocks are traversed by garnet-diopside-quartz veinlets described by Goodspeed and Coombs (1932, p. 554).

Rhyolite is scattered throughout the Naches formation, and some of it may be intrusive. The contact relationships are visible at few localities.

In hand specimen, it is aphanitic and light colored, commonly some shade of yellow or buff; however light gray, blue gray, and green rhyolites are also present. Some are massive, but others have a clastic texture shown by color differences between the fragments and the groundmass. At other places, this rhyolite is flow-banded with complex swirls.

The banded rocks contain partly resorbed quartz phenocrysts and oligoclase phenocrysts partially altered to sericite and kaolin in a devitrified groundmass. The banding is shown by slight differences in crystal size, slight color differences, and localization of sericite and chlorite. In a few rocks, the outlines of what may have been glass shards occur in this devitrified groundmass. These rocks also contain a few opaque grains and some masses of chlorite and opaque minerals that may be the alteration products of ferromagnesian minerals.

The fragmental-textured rhyolites are generally in the lapilli size range and are composed of lithic fragments, commonly of rhyolite, in a devitrified groundmass.

Most of the rhyolites are altered. In hand specimen, the more altered rhyolites are chalky and tiny spherulites are megascopically visible in many. In thin section, many of these rocks show a plexus of spherulites that masks the original textures. These spherulites seem to be formed of radially intergrown quartz and orthoclase and form a dark cross between crossed nicols. In some rocks, perlitic cracks also occur.

A few andesites and dacites also occur in the Naches formation. These rocks are similar to the rhyolites but contain only plagioclase phenocrysts.

Pyroclastic rocks are of limited extent and are scattered throughout the Naches. Poor exposures and alteration make it difficult to distinguish the pyroclastic rocks from volcanic sedimentary rocks. The criteria used was presence or absence of clasts of chert or quartz and, in some places, the degree of bedding.

The pyroclastic rocks grade from lithic tuff to lithic lapilli-tuff. They are buff, brown, or dark greenish gray and weather brown or tan. The clastic texture is obvious where weathering has etched out the less resistant fragments. The fragments are altered lava and now consist principally of chlorite, but there are also fragments of pilotaxitic lava. The groundmass is very fine-grained chloritic material. The larger fragments in the lapilli-tuff are dark maroon, dark green, and various shades of dark gray.

The volcanic sedimentary rocks constitute only a small part of the Naches formation. These rocks are commonly fine-grained and range from greenish brown to medium greenish gray. They are exposed along the west side of Little Kachess Lake. Some are distinctly bedded, but most are massive. They are composed of fragments of altered basalt mixed with angular fragments of quartz and plagioclase in a chloritic matrix. Commonly they are very poorly sorted, and many are traversed by small calcite veinlets.

Arkose is the most common sedimentary rock in the Naches. It is commonly light gray but in many places contains much carbonaceous material. It weathers to medium or dark brown. Commonly it is a massive, thick-bedded unit, so that facing is difficult to determine.

The arkose is 20–50 per cent quartz (including a little quartzite and chert), 10–25 per cent plagioclase, about 5 per cent orthoclase, 10–25 per cent chlorite and biotite altered to chlorite and opaque materials, up to 10 per cent muscovite, 5-10 per cent opaque materials (including iron ores and carbon), up to 5 per cent epidote, and a few per cent of any of the following: apatite, calcite, garnet, rutile, sphene, microcline, and granophyre. Much of the plagioclase is albite, but some is andesine. The fragments are equant and angular, and the micas are bent. In some areas, there is much carbonaceous and woody material. In some rocks, carbon lies on the bedding planes as a thin film between the sand grains.

The arkose grades into conglomerate. The conglomerate is granule and pebble conglomerate and commonly has an arkosic groundmass. The larger fragments, in order of importance, are quartz, basalt, schist, quartzite, and phyllite. The assemblage suggests that there was more than one source area. The basalt fragments may have come from the Naches basalt. The arkosic matrix probably came from a granitic terrane. The schist and phyllite fragments may have come from the Easton schist, which underlies the Naches formation to the south.

One special type of conglomerate is composed of an arkose matrix with fragments up to cobble size of fine-grained carbonaceous sandstone.

About three-quarters of a mile above the mouth of Thetis Creek is a conglomerate composed of boulders of serpentine in a serpentine matrix.

With decreasing grain size, the arkose grades into siltstone and mudstone. Commonly, as the grain

size decreases, the amount of carbonaceous material increases. Hence most of these rocks are dark gray or black. At many places, especially on Gale and Thetis creeks, exposures of these shales are extensive. Commonly the bedding planes of the thin-bedded rocks are marked by abundant detrital mica.

Keechelus andesite.[1]—The Keechelus formation was named by Smith and Calkins (1906, p. 8). They designated no type section, but most later workers, although they usually left it unstated, have considered the east shore of Lake Keechelus as the type (Goodspeed and Coombs, 1937, p. 12). The Keechelus is a thick, very widespread unit, composed of andesitic flows, tuff, and breccia, that extends at least 50 miles south of this area. It is extremely variable lithologically and is commonly altered. Work in different areas has resulted in divergent opinions regarding its origin and unity. The problem was first noted by Smith and Calkins (1906, p. 8), who suggested the possibility of an unconformity in this formation:

"The lower part of the series, comprising by far its greater portion, consists of a succession of volcanic rocks that show considerable alteration. The Keechelus rocks of the northern part of the quadrangle and most of those exposed in the southern part are of this character. In the basins of Greenwater and Naches rivers, however, there are some fresh lavas and tuffs, which are lithologically distinct from the more widely distributed decomposed rocks. . . .

"The discrimination of the younger lavas, based primarily on lithologic grounds, is supported also by structural and topographic data and one bit of doubtful stratigraphic evidence. . . .

"Lithologically, some of the lava in the Naches and Greenwater basins appears as fresh as any that might be collected from the slopes of the recently extinct volcanoes of the Cascade Range, and in the vicinity of Arch Rock there are great volumes of yellow fresh-looking tuff. The difference in alteration between the older and the younger lavas is made still more evident by microscopic study.

"As regards structure, the older portion has been somewhat tilted and gently folded, . . . while the younger rocks display only such gentle dips as might be considered the initial slope of the lava flows.

"The topographic evidence is clear in the area about Naches Pass and . . . may also be made out by inspection of the map. The summit of the ridge here is broad and flat, in contrast with the rugged topography to the north, and has every appearance of being near the original surface of the flow and but slightly modified by erosion. The lava forming this ridge appears to have been poured out after the extensive erosion which the Cascade Mountains suffered in Pliocene time. At other localities, however, lava showing a plateau-like upper surface is believed for other reasons to belong to the older portion of the series. The only stratigraphic evidence available is afforded by a locality on the ridge north of Crow Creek, where a little andesite apparently overlies the Ellensberg sandstone, but the contact is not well enough exposed to prove this relation."

From this they concluded,

"The conditions above described convinced the authors that the area mapped as Keechelus comprises lavas of two distinct ages.

"...the criteria, while sufficient to establish the presence of two distinct groups of these volcanics, fail, except locally, to serve as a basis for the determination of boundaries between them.

"... [in some] areas certain phases of the two are so similar that they cannot be discriminated with certainty, and the endeavor to map them separately was therefore abandoned."

Careful reading of these passages shows that Smith and Calkins recognized that they were including locally in the Greenwater and Naches River basins some apparently younger andesites with their Keechelus. Younger andesites are common in the Cascades, and Smith and Calkins suggested that this was the case, saying,

"Lower in the Naches Valley, in the Ellensburg quadrangle, late andesitic lavas occur and have been mapped, since there they are in contact with basalt rather than with lava of similar composition."

The name Keechelus was carried north by W. S. Smith (1915), west by R. E. Fuller, (1925, unpub. M.S. thesis, Univ. of Washington, p. 56) and south by Coombs (1935, p. 336). These extensions were made, for the most part, on lithic similarity. Coombs (1936, p. 150) observed that near the Snoqualmie granodiorite, the Keechelus changes from extremely variable to homogeneous.

Warren (1936, p. 242–247), working in the Mt. Aix quadrangle, suggested an upper and a lower Keechelus. In 1941, he separated the two, naming the upper Keechelus the Fifes Peak andesite. He also corrected Smith and Calkins' misinterpretation of the relationship of the Keechelus and Fifes Peak to the Yakima basalt, showing that both are unconformably below the Yakima (p. 799–804).

A. T. Abbott (1953 unpub. PhD thesis, Univ. of Washington, p. 44) remapped part of Warren's area in the Mt. Aix quadrangle, reiterating Warren's breakdown but changing the location of the contact between the (lower)

[1] The U.S. Geological Survey Lexicon of geologic names follows Smith and Calkins (1906) and gives the name Keechelus andesitic series. This is incorrect nomenclature for a rock unit. Keechelus andesite is used here, although Keechelus formation might be better because of its diverse composition.

Keechelus and the Fifes Peak andesite. W. N. Laval (1956, unpub. PhD thesis, Univ. of Washington, p. 118–120) compared the petrographic character of the Yakima basalt in the central Columbia Basin with the rocks described by Abbott and suggested that the Fifes Peak might be part of the Yakima basalt.

Studies along the west side of the Cascades showed other inconsistencies. Coombs (1935, p. 336) reported the Keechelus to be unconformable over the Puget formation near Mt. Rainier. Warren *et al.* (1945) reported an apparently concordant contact between these two units in an area north of Mt. Rainier. Fisher (1954, p. 1340), working south of Mt. Rainier, observed andesites he called Keechelus to be intercalated with the Puget formation.

Smith and Calkins had thought the Keechelus was Miocene because they believed it to be above Guye and Yakima, both of which they dated as Miocene. Warren (1941) showed that the Keechelus is pre-Yakima. The Guye is now believed to be Eocene. Grant (1941, p. 590) reported an oreodont in the Keechelus in the Mt. Aix quadrangle, which suggests an Oligocene-Miocene age for the Keechelus.

The Keechelus probably had more than one source, especially in view of the large area covered by this formation. If so, lavas from different centers of eruption may have had different initial dips. If these lavas came into contact, the relationships could be mistaken for an unconformity. This may explain some of the reported unconformities in the Keechelus. Detailed mapping over a broad area will be necessary before this hypothesis can be evaluated.

In addition to the stratigraphic studies summarized, petrographic studies of the Keechelus were also conducted. Goodspeed and Coombs (1937, p. 12) reported that

> "A detailed study of new exposures of these breccias at the 'type locality' along the eastern shore of Lake Keechelus has led the writers to question the validity of the original explanation of pyroclastic origin. They offer instead the interpretation of *recrystallization replacement* to be applied to certain of these breccias. By this mechanism it is believed that a sandy shale has been gradually changed into an igneous-appearing rock as a process of additive hydrothermal metamorphism."

Goodspeed, Fuller and Coombs presented another more spectacular example of this process (1941):

> "At a number of localities in western Washington dacitic and rhyolitic rocks show gradational contacts with sedimentary rocks and are associated with replacement breccias" (p. 190).

> "Later work has added to our knowledge of the Keechelus andesitic series in the Snoqualmie region. It is known that it covers thousands of square miles in the Cascades of Washington. . . . Recent petrologic work in limited areas of the lower Keechelus warrants, in the opinion of the writers, the suggestions that some of the breccias, seemingly igneous, have resulted from the metamorphism of sedimentary rock. Numerous feldspar crystals in the sedimentary rock are definitely porphyroblasts. The concentrated developments of these crystals along fractures and in scattered patches produced a rock which has been termed a replacement breccia" (p. 191).

The area mapped in this study is usually considered the type area, and the relationship of the rocks in this area to some of the rocks discussed by the other workers is not known and is not likely to be known for some years.

OCCURRENCE AND RELATIONSHIPS: The Keechelus andesite, as used by the writer, occurs in three places in the western part of the area mapped. The largest area is Rampart Ridge, between Gold Creek and Little Kachess Lake; the others are near Silver Peak and on the summit ridge of Denny Mountain. At all these localities, the Keechelus lies with pronounced unconformity on the older rocks, and the only younger rock in contact with it is the intrusive Snoqualmie granodiorite. The problems in mapping are discussed under the Naches formation.

The largest area of Keechelus seems to contain rocks of different origins. From Rampart Ridge eastward, the Keechelus andesites are distinctly bedded, and their open synclinal structure was mapped. The rocks are somewhat altered here, and the bedding is much more distinct on the west face of Rampart Ridge than to the east. The stratification of the rocks on Rampart Ridge is easily seen from U. S. 10. The Keechelus here unconformably overlies the Naches formation. The rocks shown as Keechelus on the east shore of Lake Keechelus have no bedding and are homogeneous over large areas. The replacement breccias described by Goodspeed and Coombs (1937, p. 21–23) as hydrothermally altered sediments occur here.

At the Denny Mountain locality, the Keechelus is an andesite breccia with fragments up to several inches in diameter. It appears to overlie Guye unconformably, but the contact is covered by talus.

The Keechelus rocks near Silver Peak are similar to those on Rampart Ridge. They dip gently south and overlie the Naches formation.

TYPE AREA: The necessity for a new type area may be questioned because, although

most workers considered Lake Keechelus the type area, it is not clear whether Smith and Calkins meant this to be the type. They described the Keechelus at many localities and at one place stated (1906, p. 8):

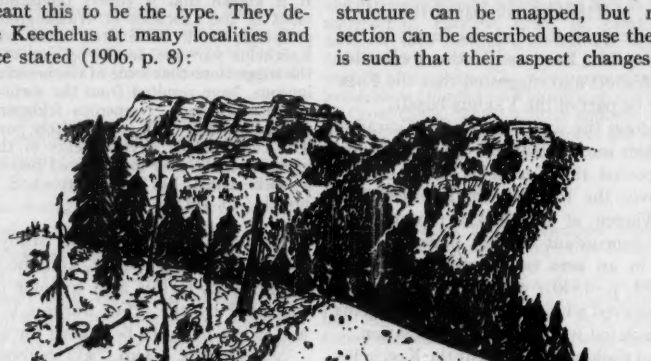

FIGURE 3.—RAMPART RIDGE, TYPE AREA OF KEECHELUS FORMATION
Looking northeast across Gold Creek Valley toward Rampart Ridge. Note glacial moraine and vegetation in foreground.

"The two divisions in their typical development, the younger as illustrated by the tuffs of Arch Rock and the lavas just south of Naches Pass, and the older as illustrated by the lavas of Pyramid Peak and the tuffs along the ridge to the north could hardly be confused."

This last statement could be considered as designating a type area, but the area is remote, and no one has used it as a type.

Before the Keechelus problem can be solved, a type area must be designated. The relationship of the Keechelus of the various workers to this type area must be established by further mapping. As noted, what is usually considered the type area for the Keechelus, along the east shore of Lake Keechelus, is metamorphosed. Therefore, it is suggested that the type area be designated as Rampart Ridge, where the Keechelus consists of a mappable thick sequence of bedded andesites. The rocks of the old type section at Lake Keechelus could not be traced any great distance.

The type area proposed herein is on the west face of Rampart Ridge at the latitude of Rachel Lake which drains into the south branch of Box Creek. In this area, a minimum of 3500 feet of section is well exposed on the west-facing cliffs (Fig. 3) and on the top of the ridge. These rocks are bedded, and their structure can be mapped, but no general section can be described because the alteration is such that their aspect changes markedly along the strike of a given bed. These rocks will be described in more detail hereafter. These rocks overlie the Naches formation unconformably and are intruded by the Snoqualmie granodiorite.

AGE: In the area mapped, the Keechelus lies unconformably above the Guye formation of questionable Paleocene or Eocene age and the Naches formation of questionable Eocene age. Thus in this area, its base may be as old as Eocene. No younger rocks are in stratigraphic contact with the Keechelus here, but it is intruded by the Snoqualmie granodiorite. The Keechelus is the youngest rock intruded by the Snoqualmie granodiorite.

The age of the Keechelus was necessarily discussed under previous work and will only be summarized here. On the basis of the oreodont found by Grant in the Mt. Aix quadrangle, the age of the Keechelus is usually considered to be upper Oligocene or lower Miocene. Waters (1955b, p. 710) summarized the relationships of the volcanic rocks of the Cascades. Although he included all the andesitic rocks in his Keechelus formation so that it overlaps in time the Eocene and the Pliocene,

he noted that "A little andesitic debris appears in the Cowlitz formation (Upper Eocene), the marine Oligocene and Miocene formations of western Washington and Oregon are flooded with it."

DESCRIPTION: The lavas are somewhat altered and so far as can be determined consist of basalt and andesite with a few rhyolitic rocks. Most of these rocks are distinctly lighter in color than most basalts. In the absence of chemical analyses, the classification is based on plagioclase composition. However, even this is difficult in practice because of differences in composition between microlites and phenocrysts, and because zoning is common in all the plagioclase crystals.

Hand specimens of the andesite range from light to dark green, although a few are dark gray. Most of these rocks have white plagioclase phenocrysts, and many also have dark-green chlorite-filled amygdules or, more rarely, white amygdules. Most of these rocks weather medium brown or red brown, but above timber line a few have a thin, light-green- to almost-white-weathering rind. They are thick-bedded to massive rocks.

In thin section, the alteration is very evident. The texture is porphyritic, commonly amygdaloidal, with a pilotaxitic or, less commonly, a trachytic groundmass composed of glass or cryptocrystalline material containing plagioclase microlites, opaque materials, and in a few rocks, pyroxene grains. Plagioclase, augite, and pseudomorphs of chlorite, epidote, antigorite, and other alteration products after ferromagnesian minerals are set in the groundmass.

The plagioclase phenocrysts range from oligoclase An_{24} to bytownite An_{73}; however, in most of these rocks, the composition is either labradorite or sodic andesine. The microlites are commonly oligoclase, but some are as calcic as labradorite. The more sodic plagioclase is more common in the altered rocks. Many of these plagioclase phenocrysts show progressive gradational zoning, but oscillatory zoning is not uncommon. The more altered plagioclase crystals are undeterminable pseudomorphs consisting of any or all of the following: kaolin, sericite, calcite, epidote, and quartz. Commonly the plagioclase phenocrysts are fractured and cloudy with inclusions of kaolin and sericite and have patches of chlorite, calcite, epidote, and quartz.

Augite is present in about half these rocks and up to 10 per cent is common in the less altered rocks. The augite is relatively fresh, but in some rocks it is altered to chlorite. There are also pseudomorphs, apparently after ferromagnesian minerals, composed of chlorite, antigorite, epidote, zoisite, calcite, and iron ores. Some of these pseudomorphs have the shape of olivine and are composed of antigorite with the rim and irregular fractures outlined by the mineraloid iddingsite.

Most amygdules are filled with chlorite, mostly radial penninite. Others are filled with quartz and more rarely with chalcedony.

Rhyolite is much less abundant than andesite and occurs as thin flows throughout the Keechelus. It ranges from almost white to a very light purple and weathers red brown. These rocks are aphanitic and some are flow-banded. On Rampart Ridge, south of Alta Mountain, there are exposures of an almost white, aphanitic rhyolite that contains tiny disseminated grains of pyrite. The ridge between the north fork of Gale Creek and Box Creek (just south of the second "E" of Wenatchee on Plate 1) is partly composed of flow-banded rhyolite. The flow banding is emphasized by fine-grained, light-purple layers. This rock is cryptocrystalline, and the only minerals that can be determined are turbid plagioclase, apparently oligoclase, and orthoclase.

Smith and Calkins described pyroxene diorite in the Keechelus volcanic rocks on Silver Peak and elsewhere and suggested on inconclusive field and petrographic evidence that these occurrences might represent the roots of the Keechelus volcanoes.

The rock on Silver Peak is a drab-green, medium-grained phanerite, in which only plagioclase and a green prismatic mineral can be seen with the hand lens. This rock is considerably altered and is composed of turbid oligoclase, a little quartz, biotite in part altered to chlorite and iron oxides, with hornblende, epidote, and chlorite as alteration products after the original pyroxene.

The pyroclastic rocks and lavas are about equal in quantity and are interbedded. The pyroclastic rocks are more altered than the lavas. These compact, dense rocks are light in color, commonly some shade of green, although a few are light purple or blue gray. They are lithic tuff, lapilli-tuff, and some breccia. They commonly weather to some shade of green or, less commonly, to brown or red brown. The texture of these rocks is well displayed in many outcrops where weathering has etched out the differences in resistance between the fragments and the groundmass.

These pyroclastic rocks are composed of fragments of altered volcanic rocks in a groundmass composed principally of fine-grained quartz and chloritic material. The lithic fragments, in general, are similar to the lavas described but are more altered. The plagioclase laths in the fragments are commonly turbid with inclusions of sericite, kaolin, and chlorite. The other fragments are composed of chlorite, epidote, and iron oxides. Amygdules are common and are usually filled with radial growths of chlorite.

A few detrital grains of quartz were noted. These may indicate that some of these rocks are volcanic sedimentary rocks.

Snoqualmie granodiorite.—The Snoqualmie granodiorite was first described by G. O. Smith and Mendenhall (1900), who recognized its Tertiary age. It was mapped by Smith and Calkins (1906, p. 9), who wrote,

"Some genetically related masses of granular igneous rocks belonging for the most part to the granodiorite type, but including more acid as well as more basic modifications, are exposed in the northern part of the quadrangle, mostly about the headwaters of Snoqualmie River, whose name is therefore adopted for the formation."

Similar Tertiary granitic rocks exposed over several hundred square miles have been described from north and west of this area (W. S. Smith, 1916; H. L. Bethel, 1951, unpub. PhD thesis, Univ. of Washington). To the south, there are several disconnected areas of Tertiary granitic rocks in the Rainier and Mt. Aix quadrangles. These rocks have been called Snoqualmie (Coombs, 1936, p. 167; Warren, 1941, p. 797; A. T. Abbott, 1953, unpub. PhD thesis, Univ. of Washington).

OCCURRENCE AND RELATIONSHIPS: The Snoqualmie granodiorite occurs around Snoqualmie Pass. A small outcrop of a similar rock, just south of Howson Creek, is shown on the map as Snoqualmie. The Snoqualmie granodiorite intrudes the Denny, Guye, Naches, and Keechelus formations. It is resistant and is carved into high peaks and ridges, such as Granite, Denny, and Snoqualmie mountains.

The contact between the Snoqualmie granodiorite and the country rock is exposed in several places. Along U. S. 10 and the Snoqualmie River, just west of Snoqualmie Pass, the granodiorite can be seen in intrusive relationship with the hornfels of the Naches formation. The hornfels zone is apparently only a local effect of the intrusion, for it disappears a few hundred feet away from the contact. On the west side of Denny Mountain the granodiorite intrudes the limestone of the Denny formation. Between Red Mountain and Gold Creek, exposures of the contact are excellent especially north of Kendall Peak where a small stream has cut deeply into the main ridge.

AGE: The youngest formation intruded by the Snoqualmie is the Keechelus which may be Oligocene or Miocene. Thus the Snoqualmie may be Oligocene or younger, as was noted, though, the dating of the bedded rocks is not secure. In the present area, the Snoqualmie is not in contact with any younger rocks, so no upper limit can be placed on its age.

Farther south, Warren (1941, p. 797) and A. T. Abbott (1953, unpub. PhD thesis, Univ. of Washington) report granitic rocks, that they called Snoqualmie, intrusive into the Keechelus but not into the Fifes Peak andesite or the Yakima basalt. They considered the younger rocks to be Miocene and dated the Snoqualmie granodiorite as Oligocene on this basis and because of the oreodont found nearby in the Keechelus. The relationship of the Keechelus and Snoqualmie there to the rocks in the area of this study is not known.

Larsen, Keevil, and Harrison (1952, p. 1050) determined the age of the Snoqualmie granodiorite using the zircon method. They reported 63 million years as the age of "granite from Snoqualmie Mountains, Washington". Because they were not satisfied with this measurement, they collected a second specimen from White River, north of Goat Island Mountain near Mt. Rainier. This specimen was run by Howard Jaffe, who obtained a good measurement of 60 million years (Larsen, written communication, February 29, 1956). Sixty million years is usually considered as Paleocene or Eocene.

DESCRIPTION: For the most part, this intrusive unit is granodiorite, but it locally includes both more basic and more felsic rocks. The Snoqualmie granodiorite and its associated dikes were described by G. O. Smith and his associates. Smith and Mendenhall (1900, p. 224) noted "There are more basic phases of the Snoqualmie granodiorite intermingled in such a way as to make their separation as futile as it would be unnatural."

In hand specimen, the granodiorite is a medium- to fine-grained, light-colored rock, composed of white feldspars, clear quartz, and prominent hornblende and biotite. The rocks, examined microscopically, range in composition from granite to quartz diorite. In all these rocks, the plagioclase is zoned, generally with andesine cores and oligoclase rims, with many individuals showing several oscillatory cycles superposed on their normal progression. Also, the feldspars are universally cloudy with sericite and kaolin, and the mafic minerals are partly altered to chlorite and epidote.

Smith and Calkins (1906, p. 9) accurately described the principal rock type as granodiorite composed of well-formed plagioclase, interstitial orthoclase, quartz anhedra, hornblende, and biotite, with apatite, zircon, magnetite, titanite, and allanite as accessories. The small intrusion near the head of Gold Creek and the other small intrusions east of Gold Creek are quartz diorite containing only limited amounts of orthoclase. The quartz diorite averages 70 per cent zoned anhedral plagioclase, 15 per cent quartz, less than 5 per cent orthoclase, 5 per cent chlorite, 5 per cent biotite, up to 5 per cent epidote (replacing the hornblende), and a few per cent of green hornblende and opaque minerals. In the vicinity of Snoqualmie Mountain, the rocks are more felsic, and the typical rock is a nearly white granite composed of quartz, altered feldspar (predominantly orthoclase), and a little biotite.

Near their contacts, many of these rocks are porphyritic with quartz phenocrysts set in a fine-grained, altered matrix.

Structure

Relationship Between the Two Sections

Table 1 shows the suggested correlation between the eastern and western stratigraphic columns of the area. As noted, the ages and correlations of these rocks are tentative, owing to the lack of distinctive fossils.

Table 1 shows that, while there is an over-all similarity between the two columns, the many local units indicate that this area was in a region of intermittant crustal unrest during Tertiary time. Several times of folding and many degrees of folding are recorded in the many unconformities. The juxtaposition of the two different columns also indicates orogeny.

North of the present area, the Paleozoic and Mesozoic rocks trend northwest. This trend is apparently old, and the Tertiary rocks of the present area follow it.

The oldest Tertiary rocks are the Swauk and Guye formations. The Swauk is a thick arkose with prominent conglomerates and appears to have been deposited under conditions that favored rapid erosion and quick burial. The Guye formation may be the same general age as the Swauk, for both are postmetamorphism and prevulcanism. The Swauk was folded along the old northwest trends, in part quite tightly. The Swauk is more intensely deformed where the basement is peridotite than where the basement is Easton schist, probably owing to the differences in response of the dissimilar basement rocks to the orogenic forces. The Mount Stuart block was the source of at least part of the Swauk sediments. Just south of the Mt. Stuart block, the northwesterly trends of the Swauk folds become more east-west, indicating that the Mt. Stuart block was an important positive structural element at the time of, and prior to, Swauk folding.

To the west, presumably at about the time of Swauk folding, the Guye sediments were folded. The Guye beds have a steep east-dipping homoclinal structure, with trends varying from north-south to northwest. This local departure from the regional northwest trends may be due to the intrusion of the Snoqualmie granodiorite.

After the folding of these rocks, the Silver Pass andesite and the Mt. Catherine rhyolite were extruded. The Silver Pass volcanic rocks and the Mt. Catherine rhyolite may be roughly contemporaneous.

The next sequence consists principally of arkose and basalt. In the western part of the area, rocks of these two types are interbedded to form the Naches formation, and to the east they are separated into Teanaway basalt and overlying Roslyn arkose. The validity of this correlation is questionable because in neither the Roslyn nor the Teanaway are sedimentary and volcanic rock intermixed. The structures bring out more differences between the Naches and the Teanaway-Roslyn. The Naches has been deformed into a series of moderately tight northwest folds. In contrast, the Teanaway-Roslyn has been, for the most part, only slightly deformed into a large basin. Although the Teanaway and Roslyn rocks have very gentle attitudes throughout most of this basin, the basin is asymmetrical; on the southwest side the beds dip steeply and on the west is a tight fold.

The Keechelus andesite was extruded on the eroded Naches. The Keechelus is preserved only in the western part of the area but may have covered all the older rocks, especially if the andesites on Goat Mountain are Keechelus as suggested by Smith and Calkins. In the western part of the present area, the Keechelus is slightly deformed into a syncline.

In the western part of the area, the Snoqualmie granodiorite was intruded sometime after the Keechelus.

In the eastern part of the area, the Yakima basalt overlies the earlier rocks unconformably. In this area, the Yakima is almost horizontal, but to the southeast, in the Yakima Canyon, it is steeply folded. The Yakima basalt was folded probably in Pliocene time when the Cascade Range was uplifted on a more or less north-south axis. The nature of this uplift is not known, but it appears to have been in part compressive (W. N. Laval, 1956, unpub. PhD thesis, Univ. of Washington, p. 143).

Kachess Fault

The main structure in the area studied is the Kachess fault which separates two different stratigraphic columns of the same general age. This north-south fault has been traced 15 miles from the crest of the Cascade Range to Kachess Lake, and it probably extends at least to the Yakima Valley, another 5 miles. At its northern end, high on the Cascade crest, the fault plane is exposed. Here the fault plane dips 65° east, and the Easton schist on the

east side has been brought into contact with Naches rocks to the west, showing that it is a reverse fault (R. C. Ellis, 1959, unpub. PhD thesis, Univ of Washington). In the area of this study, the fault plane is concealed by the Kachess lakes, but the effects of the fault are apparent. East of Little Kachess Lake, the valley sides rise 3300 feet and are composed of Easton schist. To the west, the valley sides are 2000 feet high and are composed of Naches formation.

The amount of movement on this fault is difficult to estimate because the same beds do not appear on both sides. This may indicate that the movement has been large. Possibly the minimum vertical movement is indicated by the 3300 feet of topographic relief of Easton schist plus at least 4000 feet of once overlying Swauk now removed by erosion. The possibility of horizontal movement on this fault must be considered, but its magnitude cannot be estimated.

It is difficult to date this fault. In the present area, (?) Eocene rocks are the youngest cut by it; however, to the north it is apparently terminated by the intrusive Snoqualmie granodiorite, and so is older than the Snoqualmie (R. C. Ellis, 1959, unpub. PhD thesis Univ. of Washington).

Near the north end of Kachess Lake this fault apparently branches with one fault in Kachess Lake and the other in the broad glaciated valley of Lodge Creek. Limited exposures near Lodge Creek and on Amabilis Mountain preclude mapping.

The Kachess fault separates two different, thick stratigraphic columns of the same general age. This may indicate that two different Tertiary basins have been brought into contact by the fault or that the fault marks a boundary between two basins. The formations of neither section appear to thin near the fault. This suggests that the two sections were brought together by fault movement, possibly of large magnitude. Smith and Calkins (1906) mapped several long faults with similar trends southeast of Kachess Lake, which also suggest that this is an important structure.

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CRITERIA FOR RECOGNITION OF LAHARIC BRECCIAS, SOUTHERN CASCADE MOUNTAINS, WASHINGTON

BY RICHARD V. FISHER

ABSTRACT

Comparison of properties of laharic (volcanic mudflow) and pyroclastic breccias leads to the conclusion that two described volcanic breccias near Mt. Rainier, Washington, originated as lahars. The two volcanic-breccia deposits are typical of many breccias within a thick andesitic series (Eocene-Oligocene), suggesting that rapid denudation of nearby volcanoes was an important process in the accumulation of the lower Tertiary andesitic material within the southern Cascade Mountains of Washington.

CONTENTS

TEXT

	Page
Introduction	127
Two examples of laharic deposits	129
Skate Creek laharic breccia	129
Randle laharic breccia-conglomerate	130
Origins	130
General statement	130
Sorting	130
Bedding characteristics	130
Composition and texture of large fragments	130
Rounding of fragments	130
Type of induration	131
Initial dip of breccias	131
Alteration of minerals	131
Carbonaceous fragments	131
Summary	131
References cited	132

ILLUSTRATIONS

Figure	Page
1. Index map showing location of the Skate Creek and Randle laharic breccias	129

INTRODUCTION

Andesitic rocks similar to rocks referred to as the Keechelus andesitic series (Warren, 1941; A. T. Abbott, 1953, unpub. Ph.D. thesis, Univ. of Washington) within the map area (Fig. 1) include epiclastic rocks such as volcanic breccias, conglomerates, graywackes, and shales; pyroclastic rocks, including pyroclastic breccias, lapilli tuffs, and tuffs; and flow rocks, including andesite and subordinate basalt, dacite, and rhyolite. Epiclastic volcanic rocks appear to be the most abundant rock type.

Coarse-grained volcanic clastic rocks are classified herein as follows:

Volcanic breccia
  Pyroclastic breccia
    Vulcanian breccia
    Pyroclastic flow breccia
  Epiclastic volcanic breccia
    Laharic breccia
  Volcanic conglomerate

Volcanic breccia is composed predominantly of angular volcanic fragments greater than 2 mm set in a subordinate matrix of any composition and texture, or with no matrix, or is composed of nonvolcanic fragments set in a volcanic-rock matrix (Fisher, 1958). Pyroclastic breccia is a volcanic breccia containing fragments produced by ejection of liquid and/or solid particles from a volcanic vent. Vulcanian breccia is composed of cognate accessory or accidental angular blocks produced by explosive, vertical ejection from a volcanic vent. This term is more restricted but is nearly synonymous with Wentworth and Williams' "volcanic breccia" (1932, p. 51–52), *i.e.*, a pyroclastic rock composed of angular volcanic

fragments exceeding 32 mm and solidified before aerial flight and deposition. Agglomerate is considered here to be a pyroclastic rock composed dominantly of volcanic fragments rounded by volcanic processes. These fragments may adhere at their points of contact (agglutinate) or lie loosely in a tuff matrix. Pyroclastic flow breccia is a breccia composed predominantly of angular fragments transported by "pyroclastic flows" (term used by R. E. Wilcox, personal communication) or "glowing avalanches". The term "welded tuff" is not used here since such rocks may not be welded (Gilbert, 1938; Enlows, 1955) and may be composed of fragments larger than tuff size.

According to Williams (1956) "glowing avalanches" originate in three major ways. Peléan-type "glowing avalanches" result from explosions through flanks of a dome (*nuée ardente d'explosion dirigée* of Lacroix) or by the collapse of a dome without attendant explosion (*nuée ardente d'avalanche* of Lacroix). Krakatoan-type "glowing avalanches" are initiated by vertical low-pressure explosions through craters (*nuées ardentes d'explosions vulcaniennes* of Lacroix) Fissure-type "glowing avalanches" are initiated by low-pressure upwelling of effervescing magma through fissures. This type of "glowing avalanche" is most common and develops typical textures and structures of "welded tuff" deposits.

Epiclastic volcanic breccia contains fragments produced by any type of rock fragmentation and transported by epigene geomorphic agents or by gravity transfer. Only laharic breccias are considered here, but all transitions from very viscous mudflows to "normal" stream transportation, and thus between laharic breccias and water-laid volcanic breccias and volcanic conglomerates, exist. Laharic breccia is formed by mudflow carrying, dispersing, and depositing coarse- and fine-grained volcanic particles and/or admixed nonvolcanic material. The word lahar comes from the Indonesian word for volcanic materials transported by water (Van Bemmelen, 1949, p. 191). They may originate (Anderson, 1933, p. 252–258) as a result of eruptions through a crater lake; eruptions causing the melting of ice and snow; eruptions following heavy rains; or eruptions accompanied by heavy rains. They are termed "hot mud flows" by Van Bemmelen. Cold mudflows ("normal lahars" of Van Bemmelen) result from collapse of the dam of a crater lake; heavy rains falling on unconsolidated ejecta; or rapid melting of ice and snow. In addition, mudflows may be initiated where "glowing avalanches" enter streams (Williams, 1956, p. 61–62).

Volcanic conglomerate is composed of fragments larger than 2 mm (Fisher, 1958, Table 1) and rounded by epiclastic processes. Volcanic conglomerates may be transitional with epiclastic volcanic breccia and are classified as follows: volcanic conglomerate-breccia, >50 per cent angular volcanic fragments; volcanic breccia-conglomerate, >50 per cent rounded volcanic fragments. Specific types of volcanic breccias may be named accordingly, *e.g.*, laharic breccia-conglomerate.

The origins of many types of volcanic breccias seen in the mapped area are doubtful. Flow breccias, laharic breccias, vulcanian breccias, and water-laid volcanic conglomerate-breccias are seen, but breccias which may have originated by other processes such as pyroclastic flows, mass wasting (gravity transfer), or intrusion have not been recognized.

Most volcanic breccias in the mapped area are massive, unsorted, and well lithified. Some beds are as much as 100 feet thick in a single outcrop, and only slight color changes seen from afar indicate bedding. The breccias are interbedded with andesite flows, volcanic sandstones, shale, and tuffs. Fragments of the breccias average about 2 inches in longest dimension, are mainly angular and/or subangular (some are well rounded), and commonly consist of four to five textural varieties of andesite. Textures range from prophyritic to aphanitic with vesicular and/or amygdaloidal rocks the minority, and colors are various shades of red, gray, green and may be black. Volcanic breccias composed of one or two fragment types are less common. Disseminated wood fragments are common. In some breccias the fragments grade into a usually lighter-colored green-gray fragmental matrix; more commonly, outlines are sharp. In places the breccias grade laterally and/or vertically into volcanic conglomerate or volcanic sandstone.

Microscopic examination of the breccias reveals a clastic texture with many broken, angular to rounded primary mineral grains and abundant secondary minerals set in submicroscopic, nondescript material with tiny weakly birefringent secondary mineral growths. Microscopic characters are not necessarily diagnostic of origin, since most of the rocks are highly altered.

Plagioclase appears as broken, angular fragments, although many euhedral grains are present, and ranges in abundance from 3 to 60 per cent. The feldspar ranges from oligoclase to

labradorite, but andesine is by far most abundant. Some feldspar grains are fresh; others are in various stages of alteration to kaolin, calcite, and, less commonly, epidote. Most A few scattered, water-worn, vesicular bomb-shaped fragments are present. Scattered, non-oriented carbonaceous fragments and one well-preserved spruce cone (*Picea* sp.) were observed.

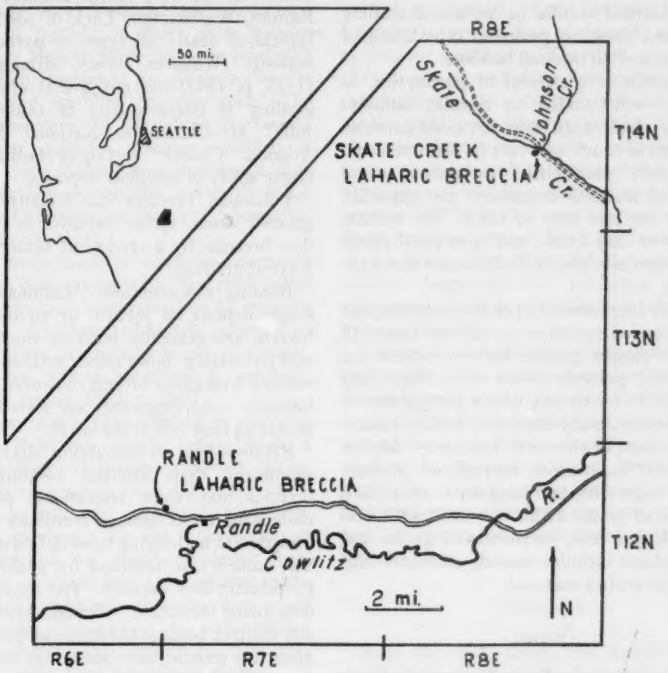

FIGURE 1.—INDEX MAP SHOWING LOCATION OF THE SKATE CREEK AND RANDLE LAHARIC DEPOSITS

augite grains are altered to chloritic material, antigorite, and magnetite. Chlorite, calcite, antigorite, and quartz are the dominant secondary minerals. Chlorite is common and presumably gives the rocks their characteristic green color.

TWO EXAMPLES OF LAHARIC DEPOSITS

Skate Creek Laharic Breccia

At the junction of Skate and Johnson creeks (Fig. 1), an andesite volcanic breccia unconformably overlies a dark-gray, pebbly volcanic sandstone layer up to 2 feet thick. The breccia is poorly sorted and without visible bedding or lamination. Fragments of the breccia are porphyritic to dense andesite, are angular and subangular (some are well rounded), and average about half an inch in longest dimension. In places the underlying pebble sandstone grades into the breccia. In thin section the matrix of the breccia contains 15 per cent angular to rounded andesite fragments and abundant secondary minerals which have altered from primary angular to rounded mineral grains. The fragments are enclosed in a submicroscopic clastic material (25 per cent) partially altered to chlorite and calcite. Grains in thin section range from submicroscopic to 4 mm. Plagioclase (20 per cent) consists of secondary growths and of remnants of primary plagioclase grains which are altered to calcite (20 per cent) and quartz (5 per cent) Kaolin (2 per cent) forms as a partial alteration product of most of the primary feldspar grains. Secondary chlorite (3 per cent) and scattered magnetite grains (0.1 per cent) may be alteration products of ferromagnesian minerals.

Randle Laharic Breccia-Conglomerate

One mile west of Randle (Fig. 1) are three well-exposed, massive volcanic breccia-conglomerate beds. The beds are poorly sorted and show no marked vertical or horizontal grading of fragment sizes, no preferred orientation of fragments, and no internal laminae.

The breccia is composed of porphyritic to aphanitic, well-rounded to angular andesite fragments which range from clay-sized particles to boulders as much as 4 feet in diameter, and disseminated, nonoriented carbonized wood chips. The andesite fragments are typically green but are also gray or black. The matrix, composed of clay-, silt-, and sand-sized particles, is irregularly distributed around the large fragments.

The rock fragments (75 per cent) are angular to well-rounded andesites in various stages of alteration. Augite grains, some of which are fresh, form 1 per cent of the rock. Plagioclase grains (An 35 to An 45), which are angular to rounded and partially altered to kaolin, constitute 6 per cent of the rock. Secondary chlorite (3 per cent) has formed throughout minerals and rock fragments, and magnetite occurs as a few scattered grains within the matrix. Calcite cement (15 per cent) surrounds all grains and in many places includes sparsely scattered submicroscopic matrix material.

Origins

General Statement.—Several criteria indicate that the Skate Creek and Randle deposits are epiclastic rather than pyroclastic. Whether they were initiated by processes mentioned by Anderson (1933), represent deposits gradational with pyroclastic flows, or result from weathering and erosion of old volcanic rocks is not known. Since the deposits are contiguous with sedimentary volcanic rocks, they probably were deposited in an area either far from centers of active volcanism or close to the center of volcanism during an inactive period. The latter seems more plausible, since thick andesite flows occur farther up and lower down in the stratigraphic section. Positive evidence of origin by glacial erosion is lacking.

The writer attempts herein to differentiate between laharic and pyroclastic breccias and to evaluate significant criteria.

Sorting.—Lack of sorting in volcanic deposits clearly indicates rapid and confused deposition and is used as one evidence of mudflow action by several authors (Anderson, 1933, p. 260; Curtis, 1954, p. 458; Rouse, 1937, p. 1281; Williams, 1949, p. 23). Crandell and Waldron (1956, p. 351), however, describe a marked decrease in the number and size of "stones" from bottom to top of a recent mudflow from Mt. Rainier, Washington. Lack of sorting is also typical of nearly all types of pyroclastic flow deposits (Williams, 1956), although Enlows (1955, p. 1245) has noted that crude vertical grading is characteristic of coarse "welded tuffs" at Chiricahua National Monument, Arizona. "Chaotic" sorting in itself is not diagnostic solely of mudflow deposits.

Vulcanian breccias can usually be distinguished from either laharic or pyroclastic flow breccias by a crude but better developed vertical grading.

Bedding characteristics.—Laminae within a single deposit of laharic or pyroclastic flow breccia are generally lacking, since mudflows and pyroclastic flows move without attendant sorting. Vulcanian breccia, however, may show laminae, since fragments are sorted according to size as they fall to the earth.

Relationships to underlying strata are more significant than internal bedding. Laharic breccias may show transitional phases with underlying rocks because mudflows are able to incorporate underlying loose debris during flow. The same is true, however, for weakly explosive pyroclastic flow deposits. The lowermost bedding plane beneath a vulcanian breccia is usually distinct because the material does not move along the ground, and bomb sags may indicate deposition from the air (Stearns and Vaksvik, 1935, p. 19–20).

Composition and texture of large fragments.—Since lahars originate in several ways, the fragments may vary considerably, depending upon the composition of the volcano, its basement rock, and the surrounding terrane. Consequently, laharic breccias are commonly heterolithologic. Large fragments of pyroclastic flow deposits, however, are domonantly of one type, either cognate essential or lithic accessory fragments, whereas accidental lithic debris is rare. Pyroclastic flows of Krakatoan type, however, tend to be heterolithologic, although many contain great quantities of pumice and other highly siliceous rock types (Williams, 1956, p. 63–64). Vulcanian breccias also are usually dominated by one or two fragments types (cognate essential and/or accessory), although lithic accidental fragments may occur.

Rounding of fragments.—In both lahars and weakly explosive pyroclastic flows, large fragments may become rounded during transport.

Perceptible rounding may occur in lahars transported long distances (Curtis, 1954, p. 458), and, since percentage of water is variable in lahars, the degree of rounding shown by fragments of lahars which have traveled the same distance may vary. Vesicular and pumiceous fragments of pyroclastic flows may become rounded very readily since they are relatively soft, but such fragments are likely to be destroyed during mudflow transportation. The presence, therefore, of abundant pumiceous and vesicular particles, whether rounded or angular, is more indicative of pyroclastic flow deposits than laharic deposits, unless the lahar originated on a pumiceous slope and transportation distance was short.

Delicate projections of vesicular fragments in vulcanian breccias may be preserved unless rock fragments have been reworked by running water or by mudflows, although presence or absence of these projections is not diagnostic, since abrasion within mudflows is generally minimal (Anderson, 1933, p. 260).

More important is the rounding of sand-size fragments. Relatively small grains have a greater tendency to become rounded during mudflow transportation than by either pyroclastic flows or vulcanian explosions. Also, there is little or no chance for the introduction of already reworked and rounded volcanic fragments (large or small) into a pyroclastic extrusion, whereas mudflows can carry any available materials. Subangular to well-rounded mineral fragments are thus more indicative of epiclastic rocks than pyroclastic rocks.

Type of induration.—Laharic, vulcanian, and pyroclastic flow breccias may all be lithified by "normal" cementation processes. Essential fragments of vulcanian breccias, however, if plastic when deposited, may adhere at their points of contact, thus forming agglomerate. Pyroclastic flow breccias may show typical welding together and flattening of fragments, although all parts of a pyroclastic flow may not be welded (Gilbert, 1938); Enlows, 1955). Welded pyroclastic flow deposits may also show distinctive columnar structure (Enlows, 1955), a feature which is absent in other clastic deposits.

Initial dip of breccias.—Mudflows and pyroclastic flows are exceedingly mobile and may carry boulders weighing many tons. The initial dip of both laharic and pyroclastic flow breccias may therefore be low. Vulcanian explosions rarely have sufficient energy to disperse large blocks and bombs far from the slopes of the parent volcano; therefore thick vulcanian breccia deposits generally have high initial dips.

Alteration of minerals.—As shown by Hay (1959), alteration of feldspar depends upon composition, amount of enclosure (a "free" crystal as opposed to one enclosed in a rock fragment), permeability of deposit, and length of time for weathering. Therefore, the presence of fresh and altered plagioclase grains of similar composition in the same rock indicates different source rocks and different lengths of time for alteration before deposition. Laharic breccias may contain minerals in various stages of alteration, whereas within a single pyroclastic layer crystals will be in the same stage of alteration since they are derived from the same magma at the same time. This and the rounding of mineral fragments are probably the best microscopic criteria to distinguish laharic from pyroclastic rocks.

Carbonaceous fragments.—Carbonaceous fragments may be mixed heterogeneously and without preferential orientation in either laharic or pyroclastic flow breccias because of lack of sorting factors in either a mudflow or pyroclastic flow. In vulcanian breccias, however, wood fragments, if caught in an explosion, will show preferred orientation of long axes parallel to bedding planes, but such occurrences are admittedly rare.

Summary

Both the Skate Creek and Randle deposits show characteristics attributable to lahars. Lack of sorting, mixing of lower contacts, large rounded fragments, and unoriented carbonized wood fragments indicate deposition by either lahars or pyroclastic flows but preclude deposition from the air. The breccias are heterogeneous in composition, lack vesicular or pumiceous fragments, and contain rounded, broken mineral fragments in various stages of alteration, thus indicating that they are laharic in origin rather than produced by pyroclastic flows. Also, there is no evidence of fumerolic action, a characteristic of pyroclastic flow deposits (Williams, 1949; 1956).

Heavy vegetation and lack of continuous exposures prevent determination of the origin of most volcanic breccias in the area, but, since most of the conspicuous volcanic breccias display characteristics similar to the Skate Creek and Randle deposits, and since water-laid volcanic conglomerates, sandstones, and shales are interbedded with the volcanic breccias

throughout much of the andesitic series, it is believed that the rocks were deposited primarily as the result of rapid denudation of large andesitic volcanic cones near or on the present site of the Cascade Mountains.

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INTRAVALLEY VARIATION IN SLOPE ANGLES RELATED TO MICROCLIMATE AND EROSIONAL ENVIRONMENT

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ABSTRACT

East- or west-trending stream valleys that have erosional slopes have been reported to be (1) asymmetrical with the north-facing slopes steeper, (2) asymmetrical with the south-facing slopes steeper, and (3) symmetrical, with both sides of equal mean slope angle. A series of measurements of angles of erosional slopes, taken in three east-trending, low-gradient valleys in the Laramie Range, Wyoming, shows that north-facing slopes there tend to be 4.42° steeper than opposed south-facing slopes; slope angle is further affected by nearness to the channel. In valleys that have channel gradients greater than 6° and greatly differing vegetation density across the valley, measurements of slope angles definitely show valley symmetry. The interpretation given is that unless the channel has been maintained against the base of the north-facing slope by greater slope wash from the south-facing slope, vegetation differences and resulting rates of slope erosion alone do not produce asymmetric valleys. Valley asymmetry that results from a variety of basic causes can be attributed to a single mechanism, asymmetric lateral corrosion by the stream.

Difference in frost action on north- and south-facing valley sides is rejected as a cause of valley asymmetry in the areas studied because (1) the asymmetry is opposite that reported in tundra regions elsewhere, and (2) the degree of asymmetry in two widely separated areas does not reflect differences in the degree of frost activity in the two areas.

CONTENTS

TEXT

	Page
Introduction and acknowledgments	133
Theoretical consideration of causes of valley asymmetry	134
Analysis of slope angles in the Laramie Mountains, Wyoming	135
Interaction of channel gradient with slope orientation in producing valley asymmetry	138
Analysis of slope angles in southern Arizona	138
Effect of erosional environment on downslope profiles	141
Summary	142
References cited	142
Appendix I. Analysis of variance of slope measurements collected in the Laramie Mountains, Wyoming	144
Appendix II. One-way analysis of variance of slope measurements taken in southern Arizona, according to life zone	144

ILLUSTRATIONS

Figure	Page
1. Two-way factorial design for a single valley	136
2. Relation between channel gradient and valley asymmetry within a single valley	141

TABLES

Table	Page
1. Mean slope measurements for each valley, arranged according to erosional environment	137
2. Power of Strahler's (1950) test for difference in slope-angle means against four alternatives	138
3. Power of paired t test	140
4. Sample parameters of Θ according to life zone	140
5. Comparison of potential frost activity in southern Arizona and southeastern Wyoming	141

INTRODUCTION AND ACKNOWLEDGMENTS

The question of whether the orientation of a valley side sufficiently affects the processes of erosion and mass transport, through the agency of differences of microclimate, to cause perceptible differences in the declivity of the steepest, straight portion of approximately mature, erosional slopes in homogeneous material has received the attention of geologists for at least 50 years. The question is inherently a statistical one; because we cannot measure

the steepness of every valley-side slope, it is necessary to make estimates of population parameters on the basis of limited sample data, and we wish to control as much as possible the accuracy of these estimates. This is best done by the methods of statistical analysis. Procedures of sampling, measurement, and analysis needed for a scientific attack on the problem of valley asymmetry are relatively simple, but only two systematic, quantitative studies have been conducted (Emery, 1947; Strahler, 1950), and the results of these two studies do not agree. No conclusive quantitative evidence has yet been adduced in support of any hypothesis.

Where it is due to differences in microclimate, the difference in angle between north- and south-facing slopes, if any, is usually small in comparison to the mean slope angle; the importance of a study of this phenomenon, then, lies in the possibility of verifying predictions of the manner of operation and effectiveness of certain gradational processes.

The conclusions that are reached below are based on a large number of slope-angle measurements taken by the writer in Arizona and Wyoming, and by Prof. A. N. Strahler in California. Their validity is limited in the strictest view to those and similar areas. However, a wide variety of rock types and climatic zones is represented in the combined sample, and the writer believes that the conclusions will prove to be generally applicable in the subhumid to arid regions of the northern hemisphere, middle latitudes. Similar information obtained in more humid regions and in the Arctic might provide a very interesting counterpart to this study.

The Department of Geology of the University of Chicago supplied the necessary support for field expenses and assistants. Dr. Wm. H. Kruskal of the Department of Statistics, University of Chicago, offered some very helpful advice in planning the sampling program. Roger Melton of Stillwater, Oklahoma, and C. J. Acker of the University of Arizona assisted in the field during the summers of 1957 and 1958, respectively. The writer is indebted to Prof. A. N. Strahler, Columbia University, and to Prof. John P. Miller, Harvard University, for critically reading the manuscript and for offering a number of helpful comments. Any errors are of course the responsibility of the writer alone.

Theoretical Consideration of Causes of Valley Asymmetry

On the question of valley asymmetry there are only three geometrical possibilities; each has been supported fairly recently in print:

(1) North-facing slopes tend to be steeper than south-facing slopes in the drier portions of the North Temperate Zone, if other factors are equal. This appears to be the consensus of geomorphologists at present and is supported in the paper by Emery (1947) and in many earlier works. (It is generally believed that the difference increases northward.) At least three plausible explanations for this can be given:

(a) Gilbert's "law of divides," states that the declivity at a point on a slope is less in proportion as the quantity of runoff over it is greater (Gilbert, 1880, p. 110). The runoff and the erosion rates on north-facing slopes are less than that on corresponding south-facing slopes because of greater vegetation density and soil development and therefore higher infiltration rate and surface resistance to the eroding stress of running water. South-facing slopes accordingly tend to be eroded more rapidly, with resulting lowered slope angle (Emery, 1947, p. 67–68).

(b) The susceptibility to erosion is greater on south-facing slopes for the reasons given above, so that a greater quantity of debris is washed down into the subjacent channel from the north, thereby shifting the channel and stream of water against the toe of the north-facing slope. The south-facing slope continues to be lowered at a greater rate than the north-facing slope, which is kept steep by being undercut frequently. The toe of the south-facing slope is buried and built up to some extent, which further reduces the over-all angle. The difference in slope angles increases until another agency, presumably rate of surface creep, enters the picture and the difference in slope angles becomes steady.

(c) In regions in which the January mean temperature is 32°F. or less, snow and ice tend to remain on north-facing slopes for longer periods than on corresponding south-facing slopes; this protects the north-facing slope and allows more rapid erosion on south-facing slopes. The result is that in regions that have the requisite temperature regime, north-facing slopes are steeper than south-facing slopes (Russell, 1931).

Of course, mechanisms (1a, b, c) are not necessarily mutually exclusive. A major

contention of this paper is that mechanism (b) has by far the greatest effect in most arid to subhumid regions where slope differences exist because of differences in microclimate across valleys.

(2) That north-facing slopes tend to be less steep than opposed south-facing slopes has been suggested by von Engeln (1942, p. 143) without restriction and more recently by Büdel (1953, p. 255) as well as by a number of earlier writers (Smith, 1949, p. 1503) for tundra and "fossil tundra" areas. As this is primarily a study of fluvially controlled regions, no attempt is made to discuss completely periglacial processes that might produce valley asymmetry. Two of the proposed mechanisms are:

(a) In periglacial regions a south-facing slope, receiving much more heat by insolation, will be frozen less frequently than the opposed north-facing slope. Consequently, corrasion of the base of the south-facing slope by the stream will be more frequent and severe and will tend to produce steeper south-facing slopes. If modern processes have not obliterated the asymmetry, it will remain as an indication of a former frigid episode in now temperate regions (Smith, 1949, p. 1503).

(b) In periglacial regions, early spring snow melt-water causes greater erosion and debris movement on north-facing slopes because frozen subsoil under the north-facing slope reduces the infiltration capacity. A greater amount of debris will be brought down into the south side of the subjacent channel, causing the stream to shift against the base of the south-facing slope and steepen it (Büdel, 1953, p. 255).

Geologists who have observed Arctic and periglacial terrain from air photographs or in the field do not agree that south-facing slopes are steeper or that these mechanisms are of greater importance than others that tend to produce steeper north-facing slopes (F. A. Melton, Don Currey, J. Corbel, personal communications). Qualitative reasoning is of little further use, and quantitative work is needed to resolve the issue.

(3) The hypothesis that north-facing slopes on the average have the same declivity as south-facing slopes is supported by data collected by Strahler in the Verdugo Hills and San Rafael Hills, California (1950, p. 806–810). On the basis of his sample of 105 slope measurements, the 99 per cent confidence interval for the true difference in mean slope angles, Δ, is

$$-1.62° \leq \Delta + 1.66°,$$

where Δ is the true mean angle of north-facing slopes minus the true mean angle of south-facing slopes, in the Verdugo and San Rafael hills. This interval includes zero and therefore supports the hypothesis of no consistent difference in mean slope angles. All slopes measured in this sample were in sharp, V-shaped ravines with narrow, steep channels, affording approximately equal opportunity for corrasion of the toe of either side of the ravine (Strahler, 1950, p. 809). Strahler found that slopes that had been protected from recent basal corrasion by an accumulation of slide-rock and other debris averaged 6.6° less steep than slopes that had recently been corraded (p. 812–813). Statistically, this difference in sample means is extremely improbable if the difference in population means does not exist.

All three situations described above could not be true in general. It is important, then, to see whether any one is true in general, or whether the particular situation in any given case depends on the existence of certain conditions of stream regimen and climate. Otherwise, the conflicting viewpoints will only serve to emphasize that much of the published thought on slope development is not based on sound facts and is of less interest to the scientist than to the historian of science.

Neither Strahler's nor Emery's work is conclusive: Strahler's sample is too small to provide a good chance of detecting actual small differences in slope angles on the order of 1.0°–1.5° (see below). The sample taken by Emery was from a region of tilted strata with a regional southerly dip; the scarp slopes facing north would introduce valley asymmetry even if differences in microclimate had no effect (Emery, 1947, p. 67).

Analysis of Slope Angles in the Laramie Mountains, Wyoming

An area in the southern part of the Ragged-top Mountain quadangle, Wyoming, was selected for study on the basis of the presence of a number of similar, east-trending valleys that have low gradients, fairly homogeneous, deeply weathered granite bedrock in which no influence of structural planes on valley-side slope angles could be discerned, distinct difference in density of vegetation on north-facing and south-facing slopes, and absence or rarity

of extensive modern gullying or channel trenching. An alluvial fill was present in each valley of third or higher order. Three separate valleys were selected for detailed slope sampling.[1] At most points on the channels, the gradients measured 0.5°–1.5°; the maximum recorded was 3.5°. In each valley, at the mouth of every extensive tributary, a small alluvial fan has been built outward into the main valley by the deposition of debris eroded during periods of runoff and left as the water is lost into the fill. These fans appear to be still growing. The region lies between 7500 and 8000 feet above sea level. When the writer visited the three valleys none contained water; the streams are presumably intermittent or ephemeral.

The mean percentage of bare-soil area on north-facing slopes was 26 per cent; on south-facing slopes it was 41 per cent; this is a difference of means of 15 per cent. Measurements of slope angles were taken at points along segments of each valley such that channel gradient, width of the main valley, and the mean slope angle appeared approximately uniform from end to end.

After some preliminary work, the experimental design for a single valley was chosen (Fig. 1) to allow consideration of the effects of the alluvial fans at the mouths of tributaries on nearby valley-side slope angles. In Figure 1 and Table 1, F_o stands for absence of alluvial fans in the neighborhood of the slope, F_f stands for facing an alluvial fan, *i.e.*, opposite a tributary entering from the other side; F_a stands for above an alluvial fan, *i.e.*, on the same side of the main channel as an entering tributary and close enough to the junction so that the toe of the slope intersects part of the fan. The six situations are mutually exclusive, and each determines what is called here the erosional environment of the slopes. From four to six slope measurements in each of the six situations were made per valley, and the average is recorded in Table 1 in the appropriate cell. Thus, a three-way factorial design was obtained; each valley is considered a "block."

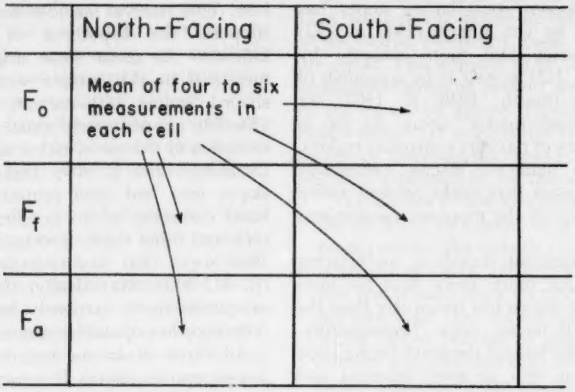

FIGURE 1.—TWO-WAY FACTORIAL DESIGN FOR A SINGLE VALLEY

The analysis of variance of the data in Table 1 is presented in full in Appendix I. Conclusions supported by the analysis are:

(1) No first-order interaction is significantly different from zero at the $\alpha = .05$ level of significance; we may assume that the quantitative effects of orientation and erosional environment within each valley are essentially additive. That this is true could be determined only by quantitative methods (Appendix IA).

(2) The column effects differ significantly from zero; north-facing slopes average $2 \times 2.2\bar{1}° = 4.4\bar{2}°$ steeper than south-facing slopes (Appendix IB).

(3) The slope angles at a distance from any tributary alluvial fan, *i.e.*, in the F_o erosional

[1] The location of the mouth of each of the three valleys is:
(1) NW1/4 sec. 21, T. 15 N., R. 71 W.
(2) SW1/4 sec. 16, T. 15 N., R. 71 W.
(3) W1/2 sec. 6, T. 15 N., R. 71 W.

environment, tend to be the same as the average of angles in the F_f and F_a erosional environments. The important conclusion is that the slope angles above an alluvial surfaces but in addition display the effects of the proximity and activity of nearby channels, that is, their erosional environment. Therefore, generalizations about slope processes and

TABLE 1.—MEAN SLOPE MEASUREMENTS FOR EACH VALLEY, ARRANGED ACCORDING TO EROSIONAL ENVIRONMENT

	North-facing	South-facing	Totals ($T_{i..}$)	Means ($\bar{X}_{i..}$)	Effects (A_i)
F_o	1. 19.12 2. 18.38 3. 21.00	1. 14.75 2. 12.38 3. 15.33	100.96	16.83	+0.01 (=16.83 − 16.82)
F_f	1. 19.33 2. 20.25 3. 23.17	1. 16.44 2. 15.50 3. 16.33	111.07	18.51	+1.69 (= 18.51 − 16.82)
F_a	1. 16.88 2. 14.50 3. 18.62	1. 13.50 2. 13.25 3. 14.00	90.74	15.12	−1.70 (= 15.12 − 16.82)
$T_{.j.}$	171.29	131.48	302.77		
$\bar{X}_{.j.}$	19.03	14.61		16.82 (Grand mean)	
B_j	+2.21 (= 19.03 − 16.82)	−2.21 (= 14.61 − 16.82)			

Blocks

Totals	Means	Effects
$T_{..1}$ = 100.06	$\bar{X}_{..1}$ = 16.68	C_1 = −0.14 (= 16.68 − 16.82)
$T_{..2}$ = 94.25	$\bar{X}_{..2}$ = 15.71	C_2 = −1.11 (= 15.71 − 16.82)
$T_{..3}$ = 108.46	$\bar{X}_{..3}$ = 18.08	C_3 = +1.26 (= 18.08 − 16.82)

fan are reduced by the same amount that slope angles facing an alluvial fan are steepened (Appendix IB).

(4) The means of all slopes within each of these particular valleys differ significantly, although the greatest difference, between valley 2 and valley 3, is less than 2.5°. This tells something about the power of this analysis to detect differences in slope-angle means.

These results reaffirm the observations of a number of workers in geomorphology and clearly imply that valley-side slopes are sensitive indicators not only of the intensity of gradational processes operating on their evolution must include considerations of erosional environment as well as of climate, vegetation, and rock and soil type. The interpretive geomorphologist must also consider possible past erosional environments as well as processes no longer active on the surfaces of slopes. A number of recent discussions of slope development have ignored this seemingly obvious point, as did the majority of earlier works.

In the present case, the high degree of association of slope angles with specific erosional environments (Appendix IB) is evidence against the hypothesis that frost action, ancient or modern, is primarily responsible for

the asymmetry of the valleys in the region studied.

Interaction of Channel Gradient with Slope Orientation in Producing Valley Asymmetry

In view of the statistical evidence for the existence of valley asymmetry presented above, why was no asymmetry detected in the statistical study conducted by Strahler? One possibility is that Strahler's sample contained too few measurements to provide an adequate chance of detecting a real, though small, difference in slope-angle means. The power of a t test to detect several alternative values of the true $\bar{\theta}_N - \bar{\theta}_S = \Delta$, based on Strahler's data (1950, p. 810), is presented in Table 2, assuming that $\alpha = .05$ and d.f. = 103 (Pearson and Hartley, 1951, p. 115). The probability of detecting a true difference of 3.0° is very good, but a difference of 0.5° or 1.5° might be undetected.

Possibly no difference in slope-angle means for north- and south-facing slopes exists in the Verdugo and San Rafael hills, in spite of the considerable difference in vegetal cover across the valleys (Strahler, 1950, p. 807). The conditions described seem ideal for the operation of agencies (1a) or (1b). However, the channel gradients are generally rather steep in the portion of the valleys having narrow channels cut in bedrock (Strahler, 1950, p. 802, map). The slopes are about four times steeper than in the Laramie Mountains, and the region is much less subject to frost action. Further work in regions that combine some of these properties was necessary.

The writer developed the following working hypothesis to explain why in some cases no difference in opposed slope angles exists, where a considerable difference in microclimate and vegetation density does exist: a steep channel gradient in bedrock allows the same runoff that brings detritus into the channel to sweep the channel clean. Such material is carried immediately out of the steep segment of the valley, in some cases as a mudflow, and is deposited either beyond the mouth of the basin on an alluvial fan or in the channel where the gradient is sufficiently lower and water is absorbed or evaporated. This prevents the accumulation of an asymmetric valley fill in the steep segment that would act to shift the channel against the base of the north-facing slope, thereby corrading it but not the base of the south-facing slope. As nothing in this would prevent the modification of the valley-side slopes by differing runoff and slope erosion rates alone (1a or c), if no difference in angles is found, then this agency acts too feebly to produce an effect. Thus, the factor of the channel gradient itself, which should be symmetric in its effect on the adjacent slopes, hypothetically would interact with and nullify the factor of slope orientation when the gradient is steep.

Table 2.—Power of Strahler's (1950) Test for Difference in Slope-Angle Means Against Four Alternatives

$\lvert\Delta^\circ\rvert$	$\phi = \frac{1}{2}\Delta\sqrt{n/S_p}$	Power
0.5	0.393	.12
1.0	0.786	.34
2.0	1.572	.62
3.0	2.36	.92

Analysis of Slope Angles in Southern Arizona

As a check on the argument presented above, and to eliminate a possible effect of slope angle on asymmetry, the writer conducted a program of sampling in valleys that have mean slope angles that range from approximately the same as those in the Laramie Mountains area to somewhat steeper but have much steeper channel gradients, roughly comparable with those in the Verdugo and San Rafael hills. Eight areas, in the Lower Sonoran, Upper Sonoran, and Canadian climatic zones were selected as follows: Care was taken to avoid sources of valley asymmetry not related to differences in microclimate on the slopes. Areas that have pronounced structural planes, modern, accelerated slope erosion, winding channels, or differences in height of divides above the channel were avoided. The valleys sampled appeared to be representative of the region. In every steep-gradient valley of the southern Arizona sample, the channel contained at most only a thin bed load; no alluvial fans were present at the mouths of tributaries.

After deciding upon a valley segment to be included in the sample, the writer and assistant established points along the channel at random. Near these points the channel gradient, γ, valley-side slope angles, θ, and the percentage

of bare-soil area, b, were measured. The error in measurement of the slope angles was probably no greater than ½°, and the error in measurement of the percentage of bare-soil area was probably no greater than 5 per cent.

The topographic variable treated is $\Theta = \theta_N - \theta_S$; the θ's were measured in pairs, opposite each other across the valley. The reason for taking paired differences instead of the quotient, as did Emery (1947, p. 62), becomes apparent when we examine the various terms that make up each slope measurement:

$$\theta_N = \bar{\theta}_{reg} + e_v + e_p + e_n + e_b + \epsilon,$$
$$\theta_S = \bar{\theta}_{reg} + e_v + e_p + e_s + e_b + \epsilon,$$

where θ_N is the measured angle on north-facing slopes, θ_S is the measured angle on south-facing slopes, $\bar{\theta}_{reg}$ is the regional mean slope angle, e_v is the quantitative effect of the particular valley studied, e_p is the effect of the particular position within that valley where the slope pair was measured, e_n and e_s are the effects of facing north and facing south, respectively, e_b is any consistent observer's bias, ϵ is the random error. Some of these terms could be zero, of course. When we take the difference, $\Theta = \theta_N - \theta_S$, the common terms drop out (*i.e.*, the regional mean, any peculiarity of the valley studied, any correlation between north- and south-facing slopes because of position within the valley, and the observer's bias), leaving $\Theta = e_n - e_s + \epsilon$, which is the quantity we wish to examine. In most cases the "random" errors are not strictly random but are positively correlated for neighboring measurements taken within a short span of time; thus the ϵ term will probably be reduced somewhat by taking differences. The quotient θ_N/θ_S would still contain all the extraneous terms in a complex and unmanageable form. The frequency distribution of Θ is very nearly symmetrical, and the cumulative distribution gives essentially a straight-line plot on arithmetic probability paper; this suggests strongly that Θ is normally distributed.

In the southern Arizona sample, the difference in percentage of bare-soil area across valleys, $\bar{b}_s - \bar{b}_n$, ranging between 15 and 45 per cent, was about 25 per cent for most valleys. Channel gradients ranged from 6° to 23° with minor modes at 8° and 15°. Slope-angle measurements ranged from 15° to 37°, with the majority lying between 19° and 30°. Forty-seven pairs of slope measurements taken were used; a number were discarded because the gradient of the intervening channel was less than 6°. The slope data are summarized as follows:

$\bar{\Theta} = -0.025\bar{5}°$
$s_\theta = 1.97°$ (standard deviation of the Θ's)
$N = 47.$

A paired t test shows:

$$t = \frac{0 - (-0.025\bar{5})}{1.97/\sqrt{47 - 1}} = 0.089, \text{ d. f.} = 46.$$

The value of t is so small that it can be expected far more than 50 per cent of the time by chance alone on repeated sampling and testing in the same way, if the population mean of Θ is zero. The 95 per cent confidence interval for the true mean value of Θ is

$$-0.61° \leq \bar{\Theta} \leq +0.55°.$$

The power of this paired t test against specified alternatives to the null hypothesis, assuming that $\alpha = .05$, is given in Table 3. We may thus be fairly confident that the true difference in the valley-side slope angles in steep-gradient valleys, if not zero, is not much greater than 1.0°, whereas it might reasonably be expected to be 4° or greater on the basis of comparison with the sample from the Laramie Mountains. As most of the actual slope measurements of this sample were in the same range of values as the measurements in the Laramie Mountains, the possibility that the lack of valley asymmetry is due to steeper slopes is eliminated.

The southern Arizona sample can be broken down on the basis of gross climatology to see what effect that has on valley asymmetry. Table 4 summarizes the data according to life zones. These means appear to support in part the hypothesis of Russell (1931, p. 484) that areas where the January mean temperature is 32°F. or less have steeper north-facing slopes, because of snow accumulation (1c). We might further conclude that the asymmetry is reversed in the hot desert areas. The analysis of these data (Appendix II) shows, however, that this amount of difference among means could be expected rather frequently by chance alone where no difference among population means exists. Despite the rather small sample size and the uneven distribution of measurements among the climate classes, the power of this analysis against departures of Θ from zero of 1° is .65; thus we may be fairly confident in rejecting the hypothesis that gross climatology in the range Lower Sonoran–Canadian affects

valley asymmetry to any appreciable degree. Many field observations confirm that in valleys that have low-gradient channels in the Lower Sonoran zone, north-facing slopes are generally at least 4° or 5° steeper than opposed south-facing slopes where an asymmetric fill is present, and the difference can be greater.

TABLE 3.—POWER OF PAIRED *t* TEST

$\lvert\Delta°\rvert$	ϕ	Power
0.5	0.871	$\approx .3$
1.0	1.742	.67
2.0	3.483	$> .99$

TABLE 4.—SAMPLE PARAMETERS OF Θ ACCORDING TO LIFE ZONE

	Lower Sonoran	Upper Sonoran	Canadian
$\bar{\Theta}$	−0.37°	0.09°	0.33°
N	15	26	6
s_θ	1.98°	2.14°	1.03°

As further illustration of the effect of channel gradient on slope-angle differences, a series of paired measurements was taken in a single east-trending valley, located in the Campo Bonito quadrangle, Arizona (N½ sec. 2, T. 11 S., R. 16 E.), near the climatic transition between the Lower and Upper Sonoran life zones. The valley is incised into an alluvial fan. Values of Θ plotted against channel gradient show a gradual increase in valley asymmetry as the gradient decreases downstream (Fig. 2); Θ probably approaches zero closely for $\gamma > 6°$.

The evidence strongly favors the conclusion that unless an asymmetric valley fill is maintained, differences in vegetation across a valley, acting through differences in rates of slope erosion, will not produce differences in valley-side slope angles and thus that the erosional environment of a slope, *i.e.*, the proximity and erosional activity of the channel, if any, at its base, is of greater importance to the slope angle than the gradational processes acting on the slope surface *per se*. This is further emphasized by the observation of Strahler's (1950, p. 812–813) that debris-protected slopes average more than 6° less in angle than slopes not so protected, and by the commonly observed fact that slopes on the outside of channel bends are considerably steeper than slopes above the inside of channel bends or above straight channel segments. Asymmetry can exist even in steep-gradient valleys, but only if the base of one slope facet is corraded more strongly than that of the opposed slope. This effectively rules out the possibility that the lack of asymmetry in spite of a considerable difference of microclimate and vegetal density can be due simply to the youthful condition of the valley or to the greater rate of channel downcutting in relation to slope erosion, as was suggested by Emery (1947, p. 68), who has also noted the lack of asymmetry in steep valleys. In steep-gradient valleys the conditions are not favorable for the maintenance of a continuous asymmetric fill, which if present would cause asymmetric basal corrasion. Thus, valley asymmetry that is the result of a variety of basic causes can be attributed to a single mechanism, asymmetric basal corrasion.

Perhaps the degree of valley asymmetry observed in areas studied here is the result of factors no longer active in these areas. One possibility is that the differences in slope angles developed solely because of differences in vegetation density and of erosional intensity on the slope surfaces (1a), and that a valley fill accumulated only after a change toward a generally drier climate, which may also have caused the initiation of the tributary channels. This hypothesis does not adequately explain why asymmetry does not always exist in valleys that have steep gradients where microclimatic differences exist and is confuted by the data from the Laramie Mountains. It would require a very clumsy *ad hoc* hypothesis to explain the observed variation in slope angles if the slopes developed entirely before the tributary alluvial fans were built. Further, in both study areas, the writer noted a rough correlation between the degree of asymmetry in a valley and the relative size of the valley fill it contained. The slopes in these areas show a remarkable adjustment to present erosional environments and must have been at least modified with the inception of present conditions.

The hypothesis that the observed valley asymmetry (or lack of it) is a result of differences in frost action or length of snow cover alone during Wisconsin times is untenable for

areas similar to those examined. Were frost action the main cause, a much greater asymmetry should occur in the Laramie Mountains same. (Compare Table 1, F_e row, with Fig. 2.) The direction of asymmetry is opposite that reported in tundra regions, where frost action is

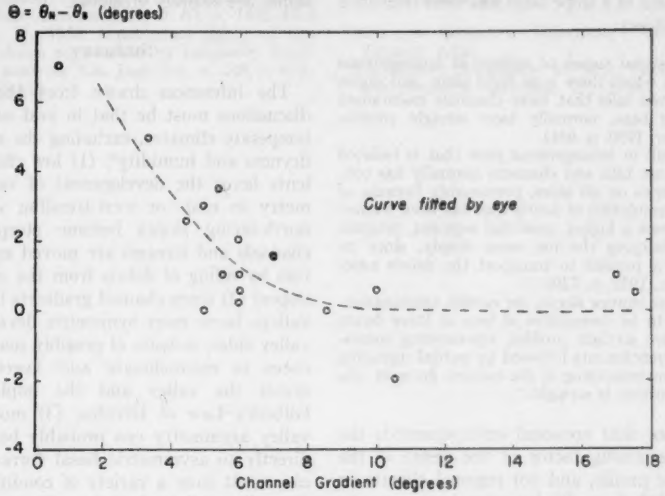

FIGURE 2.—RELATION BETWEEN CHANNEL GRADIENT AND VALLEY ASYMMETRY WITHIN A SINGLE VALLEY

TABLE 5.—COMPARISON OF POTENTIAL FROST ACTIVITY IN SOUTHERN ARIZONA AND SOUTHEASTERN WYOMING

All figures are from Visher (1954). Because of the small scale of the maps and the generality of the sources of data, these figures should be considered only as rough comparisons.

Measure of frost activity	Southern Arizona	Southeastern Wyoming
Normal annual number of days with mean temperature of 20° to 30°	0	>100
Normal annual number of days continuously below freezing	<1	≈45
Normal annual number of nights with frost	<30	≈180
Average annual number of alternations of freezing and thawing	≈25	≈130
Average depth in inches of frost penetration	<3	>20

than in the Lower Sonoran zone of southern Arizona. Table 5 compares modern frost intensity and the frequency of alternation of freeze and thaw in the two areas. The difference may have been greater during the Wisconsin. In neither area is the soil of a type to be subject to much frost heaving. In spite of the considerable difference in potential frost activity, in valleys that have low-gradient channels on the order of 1° – 2°, and away from extraneous factors such as channel bends and alluvial fans, the degree of asymmetry is nearly the morphologically important, although this fact cannot be given much weight.

EFFECT OF EROSIONAL ENVIRONMENT ON DOWNSLOPE PROFILES

The literature contains much discussion of the interpretation of straight, convex, concave, and sigmoidal downslope profiles. The following three points, which are based on purely qualitative observations of many hundreds of individual hills and valleys located in Arizona.

Utah, New Mexico, Colorado, Wyoming, South Dakota, Iowa, and Illinois, are presented to point out that the concept of the erosional environment of a slope facet has some relevance to this subject:

(1) Erosional slopes of valleys in homogeneous rocks in which there is no flood plain, and slopes of isolated hills that have channels maintained at their base, normally have straight profiles (Strahler, 1950, p. 681).
(2) A hill in homogeneous rock that is isolated from other hills and channels normally has concave slopes on all sides, presumably because of the accumulation of debris that has been washed down from a higher, erosional segment, progressively burying the toe more deeply, since no stream is present to transport the debris away (Lawson, 1932, p. 720).
(3) Most convex slopes, on careful examination, appear to be composites of two or three facets that have straight profiles, representing successive entrenchments followed by partial regrading and some smoothing of the corners. In most, the lower portion is straight.

Evidence that erosional environment is the chief determining factor of the figure of the downslope profile, and not regional climate or type of lithology (if homogeneous), can be found in many places where a hill on one side faces an open plain and on the other faces a valley and has an active stream impinging on its base. On most, the profile facing the plain is concave, and the profile facing the valley is straight. These relations are well displayed along the Black Canyon Highway (Arizona Highway 69) between Phoenix and Bumble Bee. Exceptions to these generalizations exist but are unusual in the subhumid to arid regions examined, and they can be explained by special conditions of rock structure or changes in stream regimen.

Sauer (1930, p. 368) has reported the existence of isolated hills in the Chiricahua area of Arizona that have concave, denudational slopes from summit to base. The concavity of such hillslopes cannot be the result of accumulation of debris near the base, as bedrock is present below a very thin cover of slide-rock. The writer examined the "hills of circumdenudation" mentioned by Sauer and many others that are similar and invariably found that an inhomogeneity in lithology existed; *i.e.*, at the summit of each hill there was a layer of rock, more resistant to weathering and erosion than the rock lower down, that retarded erosion of the summit relative to the lower slopes, thus producing the concavity. This relation appears to be so invariant that the writer believes it can be used with confidence to predict inhomogeneities that are not discernible from a distance, provided the hill slopes are entirely erosional.

Summary

The inferences drawn from the foregoing discussions must be that in arid to subhumid temperate climates, excluding the extremes in dryness and humidity[2], (1) low channel gradients favor the development of valley asymmetry in east- or west-trending valleys, the north-facing slopes become steeper as the channels and streams are moved against their toes by filling of debris from the south-facing slopes; (2) steep channel gradients in V-shaped valleys favor more symmetric development of valley sides, in spite of possibly marked differences in microclimate and vegetal density across the valley and the implications of Gilbert's Law of Divides; (3) most cases of valley asymmetry can probably be attributed directly to asymmetric basal corrasion, which can result from a variety of conditions within the valley; (4) both the maximum slope angle and the figure of the downslope profile are functions of the erosional environment of the slope, although the latter is strongly influenced also by rock inhomogeneities; (5) the effects on slope angle of several components of the erosional environment of the slope are additive.

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[2] Excluding the dry extreme because no vegetation will grow on either slope; excluding the humid extreme both because no data from such areas were included, and because presumably sufficient moisture is available on all slopes for a complete vegetal cover. Further work is needed in these areas.

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APPENDIX I.—ANALYSIS OF VARIANCE OF SLOPE MEASUREMENTS COLLECTED IN THE LARAMIE MOUNTAINS, WYOMING

A. Analysis of Variance—Test for Interactions

Source of variation	Sums of squares	Degrees of freedom	Mean squares	F
Row effects				
1. $\frac{1}{2}(F_f + F_a) - F_o$	.00034	1	.00034	
2. $F_f - F_a$	34.44	1	34.44	
Column effects	88.05	1	88.05	
Blocks effects	17.01	2	8.50	
Row × column interaction	4.18	2	2.09	2.21 (2, 4)*
Row × block interaction	1.23	4	0.31	0.33 (4, 4)†
Column × block interaction	3.82	2	1.91	2.02 (2, 4)*
Residual	3.78	4	0.945	
Total	152.51	17	.....	.....

B. Analysis of Variance—Test for Main Effects

Source of variation	Sums of squares	Degrees of freedom	Mean squares	F
Row effects				
1. $\frac{1}{2}(F_f + F_a) - F_o$	.00034	1	.00034	.00031 (1, 12) (n s)
2. $F_f - F_a$	34.44	1	34.44	31.89 (1, 12) (.001)
Column effects	88.05	1	88.05	81.53 (1, 12) (.001)
Block effects	17.01	2	8.50	7.87 (2, 12) (.01)
Pooled	13.01	12	1.08	
Total	152.51	17	.....	.....

* $F_{.05}(2, 4) = 6.94$
† $F_{.05}(4, 4) = 6.39$

APPENDIX II.—ONE-WAY ANALYSIS OF VARIANCE OF SLOPE MEASUREMENTS TAKEN IN SOUTHERN ARIZONA, ACCORDING TO LIFE ZONE

Analysis of Variance

Source of variation	Sums of squares	Degrees of freedom	Mean squares	F
Among means	2.86	2	1.43	0.36 (2, 44)*
Within groups	174.95	44	3.98	
Totals	177.81	46	.....	.....

* $F_{.05}(2, 44) = 3.21$

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CARBONATE REPLACEMENT OF DETRITAL CRYSTALLINE SILICATE MINERALS AS A SOURCE OF AUTHIGENIC SILICA IN SEDIMENTARY ROCKS

BY THEODORE R. WALKER

ABSTRACT

Petrographic studies show that carbonate replacement of detrital crystalline silicate minerals is an important and widespread process in sedimentary rocks. Investigations of three different rock suites indicate that authigenic siliceous features commonly occur in or near strata that contain partially to completely replaced silicate grains. This association implies a genetic relationship and suggests that silica released by carbonate replacement may be an important source of authigenic silica in some sedimentary rocks.

Such replacement is probably more common in sedimentary rocks than has been suspected because completely replaced grains commonly leave no evidence of their original presence in the sediment, or the replaced grains are preserved as carbonate pseudomorphs that can be mistaken for grains of clastic carbonate. Even in occurrences containing relicts of incompletely replaced silicate grains it generally is not possible to determine to what extent replacement has taken place.

The mechanism of solution, migration, and reprecipitation of silica is uncertain. Earlier writers have suggested that carbonate replacement of opal and subsequent reprecipitation of the released silica occurs in response to pH variations. If so, the replacement and reprecipitation of silica from crystalline silicates might be similarly explained. Recent data, however, cast doubt on this theory.

CONTENTS

TEXT

	Page
Introduction	145
Acknowledgments	146
Replacement features	146
General statement	146
Features of pseudomorphic replacement	146
Features of nonpseudomorphic replacement	146
Features of selective replacement	147
Examples of authigenic silica derived from replacement sources	147
Mechanism of replacement and reprecipitation of silica	147
General statement	147
Time of replacement	148
Distance of silica migration	149
Cause of replacement	149
References cited	151

ILLUSTRATIONS

Figure	Page
1. Solubility of amorphous silica and calcium carbonate in waters of varying pH (from Correns, 1950)	150
2. Solubility of amorphous silica in waters of varying pH (from Alexander, Heston, and Iler, 1954)	150

Plate	Facing page
1. Carbonate replacement of crystalline silicate minerals	148
2. Carbonate replacement of crystalline silicate minerals	149

INTRODUCTION

A fundamental problem in studies of the origin of commonplace authigenic siliceous features such as replacement chert, silicified fossils, and silica cement is the source of silica. Many articles deal with the origin of these features, and several potential sources of authigenic silica have been suggested. These sources are summarized by Pettijohn (1957, p. 440–444; 656–659) and Siever (1957).

Little attention has been paid in geologic literature to that silica released by carbonate replacement of detrital crystalline silicate minerals. Many investigators have observed that such minerals commonly are replaced by

interstitial carbonate, but to the writer's knowledge none has discussed the importance of this process as a mechanism for providing authigenic silica in associated sedimentary rocks. The writer first noted the significance of this phenomenon as a silica source while investigating the effect of carbonate replacement of quartz as a cause of grain frosting (Walker, 1957). Subsequent studies have revealed that replaced silicate grains commonly occur in or near strata containing authigenic silica, and the association strongly suggests that the silica is derived at least in part from that released by replacement. This paper illustrates characteristic features of such replacement and discusses three occurrences of authigenic silica believed to have been derived at least partially from this source. In addition, some general statements are made concerning the replacement mechanism.

The hypothesis suggested here is modified from Correns (1950) and Newell *et al.* (1953, p. 162–165), who have suggested that authigenic silica commonly originates from carbonate replacement of amorphous silica (opal) in tests of silica-precipitating organisms such as sponges, radiolaria, and diatoms. The expansion of that hypothesis in this paper to include replacement of more commonly occurring crystalline silicate minerals greatly increases the importance of carbonate replacement of silicates as a source of authigenic silica.

Acknowledgments

The writer gratefully acknowledges a grant from the Council on Research and Creative Work of the University of Colorado. The cost of collecting some of the data presented here was financed by the Pan American Oil Corporation from funds granted to the author for a study now in progress of silicification in the Minturn formation in central Colorado. Special thanks also are expressed to E. E. Wahlstrom and W. C. Bradley for their many valuable comments.

Replacement Features

General Statement

Most of the common silicate minerals in sedimentary rocks appear to be susceptible to replacement by carbonates, and the features produced are similar to many of those described for replacement ore minerals (Edwards, 1947, p. 94–108; Bastin, 1950, p. 33–55). Some remarkable examples of carbonate replacement of glauconite, quartz, and feldspar are shown in Plates 1 and 2. These illustrate some of the characteristic replacement features that have been observed by the writer.

Features of Pseudomorphic Replacement

Pseudomorphs form when the original outlines of detrital silicate grains survive replacement either because the boundaries are marked by concentrations of inclusions or because the texture of the replacing carbonate differs markedly from that of surrounding carbonate (Pl. 1).

Figures 1 and 2 of Plate 1 show stages in the formation of calcite pseudomorphs of glauconite grains in the Lion Mountain member of the Riley formation (Cambrian) in central Texas. Figure 3 of Plate 1 shows a pseudomorph of a euhedral overgrowth on a quartz grain in the Oneota formation (Ordovician) near Madison, Wisconsin.

Although pseudomorphs are reliable criteria of replacement, they are not always readily identified in petrographic studies. Completely formed pseudomorphs, for example, may be mistaken for grains of clastic carbonate and hence may escape recognition unless stages of replacement are revealed by incompletely formed pseudomorphs. Figure 4 of Plate 1 shows stages in the formation of pseudomorphs that resemble clastic carbonate grains in the Hygiene sandstone member of the Pierre formation (Cretaceous) near Boulder, Colorado. The original outlines of the plagioclase grains (grains A and B) are clearly preserved, and these grains are obviously incomplete pseudomorphs. Grains C and D, on the other hand, are composed entirely of calcite in the plane of the thin section, and although they appear to be grains of clastic carbonate, their proximity to grains showing various stages of replacement suggests that these grains may be complete pseudomorphs. Possible pseudomorphs of similar origin in clastic limestone are illustrated in Figure 5 of Plate 1.

Features of Nonpseudomorphic Replacement

Original outlines of replaced grains commonly are not discernible because they are not marked by inclusions, and the texture of the replacing

carbonate merges with that of interstitial carbonate. Such replacement is detectable only where relicts of partially replaced silicate grains remain in the rock. Figure 1 of Plate 2 shows this type of replacement in a feldspar grain in a calcarenite bed in the Dakota formation (Cretaceous) near Loveland, Colorado. If the replacement shown here was complete, there would be no evidence that feldspar grains were part of the original clastic detritus in this rock. Furthermore, the advanced stage of replacement suggests that if smaller feldspar grains originally were present, they probably have been replaced completely. In occurrences like this it commonly is not possible to determine the extent of replacement.

Features of Selective Replacement

Replacement is selective in most occurrences, certain mineral species being more readily replaced than others. Feldspar, for example, generally is replaced more readily than quartz (Pl. 2, fig. 2). Even within a single mineral species, certain grains or zones within grains are preferentially replaced because of subtle differences in crystal structure or chemical composition or both. Overgrowths on detrital quartz grains, for example, tend to be replaced more readily than the quartz in the core (Pl. 1, fig. 3). In addition, replacement commonly is localized along cleavage directions (Pl. 1, fig. 4) or follows incipient fractures (Pl. 2, figs. 5, 6).

Examples of Authigenic Silica Derived from Replacement Sources

The writer has observed in petrographic studies of a variety of siliceous rocks that replaced silicate grains and authigenic siliceous features commonly occur in the same strata. Although a specific source of authigenic silica cannot definitely be established in any of these occurrences, the association of replaced silicate grains in the same rock with authigenic silica suggests a genetic relationship and has led the writer to conclude that silica dissolved during replacement provides a significant source of that precipitated authigenically. The following three examples selected from those discussed heretofore illustrate this point.

(1) The Lion Mountain sandstone (Cambrian) in central Texas is highly glauconitic and contains lenses and beds of glauconitic limestone. Much of the glauconite in these rocks has been extensively replaced by calcite (Pl. 1, figs. 1, 2). The examples illustrated were collected at a locality where local concentrations of authigenic iron oxide have formed abundant hematite concretions, and authigenic overgrowths occur on most of the detrital quartz grains. Inasmuch as glauconite is composed predominantly of iron and silica, the association of replaced glauconite grains, authigenic iron oxide, and authigenic quartz in this rock is probably not fortuitous, but has resulted at least in part from local reprecipitation of iron and silica released by replacement.

(2) The Hygiene sandstone (Cretaceous) in central Colorado locally contains abundant authigenic silica as overgrowths on detrital quartz grains (Pl. 1, fig. 4). It also contains abundant detrital plagioclase grains that show various stages of replacement similar to those illustrated. In view of this association it seems reasonable that much of the silica released by replacement of plagioclase grains has reprecipitated locally as quartz overgrowths. If this is not true, one is faced with the problem of explaining what has happened to silica released by replacement of plagioclase as well as accounting for some other source of silica in the quartz overgrowths.

(3) The Minturn formation (Pennsylvanian) in the McCoy area, central Colorado, consists predominantly of calcareous arkose interbedded with limestones that commonly contain abundant detrital quartz and feldspar grains. Authigenic silica is present in many of the limestones as replacement chert bodies and silicified fossils. Thin-section studies show that detrital quartz and feldspar grains in the limestones and arkoses commonly are extensively replaced by calcite (Pl. 2, figs. 2, 3, 4). In addition, siliceous sponge spicules also showing extensive replacement are locally abundant in the limestones. Thus it appears that silica originally present in this rock in both amorphous and crystalline silicate grains has been released to interstitial waters by carbonate replacement and redistributed through the rock.

Mechanism of Replacement and Reprecipitation of Silica

General Statement

The examples given provide compelling evidence that carbonate replacement of crystalline silicate minerals is important in postdepositional redistribution of silica in sedimentary rocks. The physical and chemical

conditions that cause replacement and reprecipitation of silica, however, are not well understood, and as new data appear the following interpretations doubtless will need modification.

Time of Replacement

Carbonate replacement of silicate grains probably can occur any time in the history of a sedimentary rock, but specific occurrences can rarely be dated with certainty. Perhaps most replacement occurs soon after deposition while host sediments become adjusted to the chemical environment at the site of deposition. Early replacement doubtless would be aided by relatively high interstitial permeability prior to cementation.

Replacement, however, also can occur during the later history of rocks, and if initial permeability is sealed by cementation, secondary zones of permeability, such as fractures and solution openings, serve as principal avenues for the replacing solutions. Evidence of carbonate replacement of quartz along secondary openings is revealed by the occurrence of euhedral dolomite crystals along fractures in chert (Pl. 2, figs. 5, 6). The original fracture in the example illustrated is clearly visible when the thin section is viewed in plain light. Dolomite fills the fracture, and dolomite euhedra, unquestionably of replacement origin, extend into the surrounding chert. Isolated dolomite metacrysts surrounded by chert occur near the fracture. The date of replacement in this example is uncertain, but obviously replacement occurred after the chert was sufficiently indurated to permit fracturing.

Replacement of silicate minerals can occur even during surface weathering of the host rock. Eckel (1938, p. 47) has reported near-surface replacement of glauconite by siderite in the Weches greensand (Eocene) in eastern Texas. Replacement there is controlled by the position of the water table. Swineford, Leonard, and Frye (1958, p. 109) also have noted replacement that is related to surface weathering. They report evidence of extensive replacement of quartz and feldspar by calcite in caliche soils that are developed on the Ogallala formation in the High Plains region.

PLATE 1.—CARBONATE REPLACEMENT OF CRYSTALLINE SILICATE MINERALS

FIGURES 1, 2.—Carbonate replacement of glauconite grains. Lion Mountain member of Riley formation (Cambrian), central Texas. Plain light, 50×. Figure 1 shows glauconite grains (dark gray) that are incompletely replaced by calcite (light gray). Original borders of glauconite grains are marked by acicular inclusions of authigenic iron oxide (black). Replacing calcite is in optical continuity with interstitial calcite cement, and cleavage planes cross grain borders (upper left). Figure 2 shows nearly complete replacement of glauconite (G) by calcite (C). Authigenic iron oxide (black) is more abundant, and quartz grain (Q) has redeveloped crystal faces. Calcite forms interstitial cement.

FIGURE 3.—Carbonate pseudomorph of a redeveloped quartz crystal. Oneota dolomite (Ordovician), Madison, Wisconsin. Plain light, 60×. The quartz nucleus of the oölite originally was surrounded by a euhedral quartz overgrowth, the outline of which is still visible at the inner margin of the oölite shell. Subsequent selective replacement of overgrowth quartz has produced a carbonate pseudomorph of the redeveloped crystal (light-gray mosaic within oölite), and replacement has tended to halt at the boundary of the original quartz core. Interstitial carbonate surrounding the oölite is dolomite.

FIGURE 4.—Carbonate pseudomorphs of detrital plagioclase grains. Hygiene sandstone (Cretaceous), Boulder, Colorado. Plain light, 80×. Grains A and B are partial pseudomorphs showing advanced stages of calcite (shades of gray) replacement of plagioclase (white patches within grains, as at arrow in grain B). Grain A shows selective replacement along cleavage planes. Grains C and D are entirely calcite and are interpreted as complete pseudomorphs of original plagioclase. Remaining sand grains are predominantly quartz, but potassium feldspar also is present. Authigenic silica is present as overgrowths on quartz grains (grain E).

FIGURE 5.—Partial and possibly complete carbonate pseudomorphs of detrital silicate grains in clastic limestone. Jacque Mountain formation (Pennsylvanian), McCoy area, Eagle County, Colorado. Plain light, 50×. Outlined grain in upper center is a partial pseudomorph showing advanced stage of calcite (light gray) replacement of a detrital silicate grain (white patches within grain). Relicts of silicate grain are optically continuous. Other outlined grains are composed entirely of calcite in the plane of the thin section and may represent complete pseudomorphs. Collected by L. E. Hatfield.

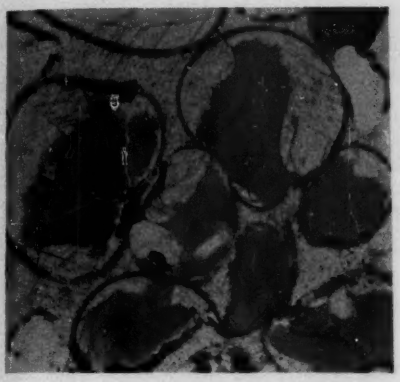

FIGURE 1

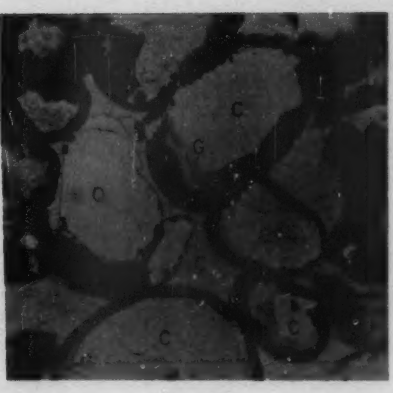

FIGURE 2

FIGURE 3

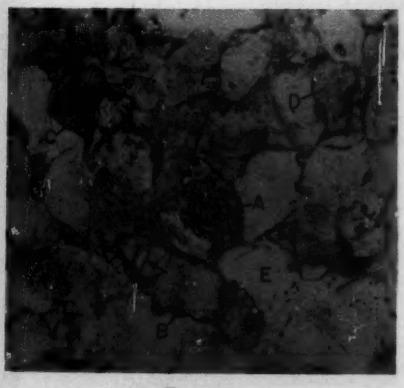

FIGURE 4

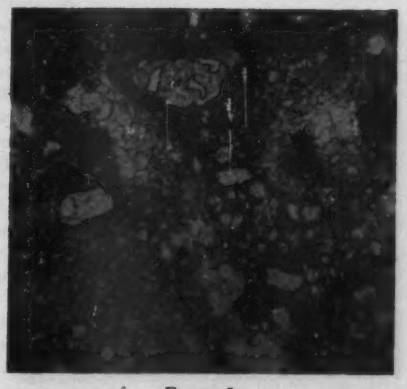

FIGURE 5

CARBONATE REPLACEMENT OF CRYSTALLINE SILICATE MINERALS

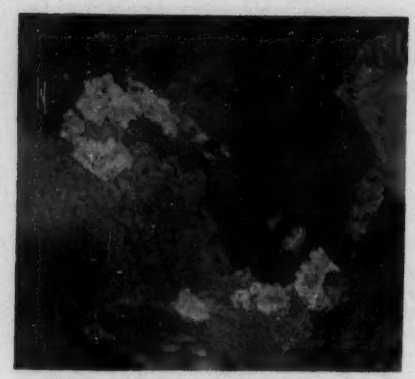

FIGURE 1

FIGURE 2

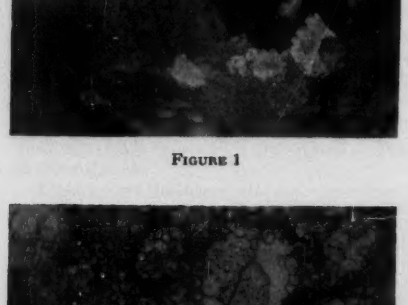

FIGURE 3

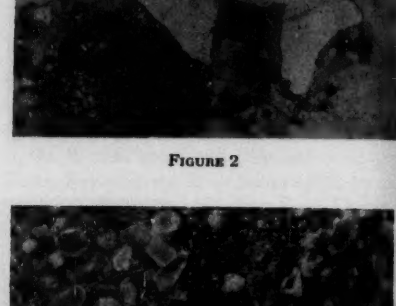

FIGURE 4

FIGURE 5

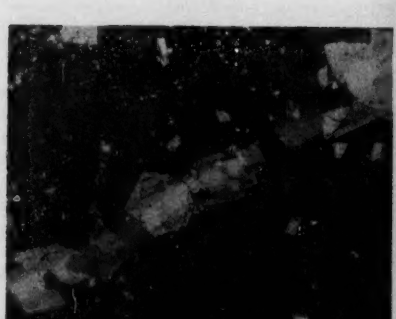

FIGURE 6

CARBONATE REPLACEMENT OF CRYSTALLINE SILICATE MINERALS

Distance of Silica Migration

The distance that dissolved silica migrates through sediments depends upon variable factors such as the rate of migration of interstitial water and the physical-chemical properties of the interstitial water and the host sediment. There is no limit to the maximum distance that silica can migrate, but authigenic siliceous features commonly occur in sediments containing replaced silicate grains, suggesting that much of the released silica does not migrate far before it is reprecipitated.

Some silica may be reprecipitated within a microscopic distance from its source, as suggested by the common occurrence of authigenic silica and replaced silicate grains in the same thin section. Figure 4 of Plate 2 shows recrystallized oölites that contain nuclei of extensively replaced detrital silicate grains surrounded by partially silicified oölite shells. This relationship suggests that silica dissolved during replacement may have migrated only as far as the surrounding oölite shells before it encountered conditions favorable for reprecipitation.

Authigenic silica in such occurrences, however, is not necessarily derived from replacement of nearby silicate grains. The conditions that favor carbonate replacement of silicate grains probably occur within relatively local areas in sediments and doubtless change with time, so that conditions favoring solution of silica at one time might later be superseded in the same place by conditions favoring silica precipitation. In this manner authigenic silica occurring in the vicinity of replaced silicate grains may originate from silica released by carbonate replacement, as well as by other processes, occurring elsewhere in the rock.

Cause of Replacement

Investigations of possible causes of replacement, migration, and reprecipitation of silica are too limited at present to be conclusive. Correns (1950, p. 53) and Newell *et al.* (1953, p. 163) have suggested that carbonate replacement of opaline silica occurs in response to pH variations in interstitial waters. If this mechanism adequately explains replacement of opal by carbonates, it probably also explains replacement of other silicate grains.

According to data published by Correns (1950) amorphous silica increases in solubility with increase in pH throughout the geologically important range between pH 5 and pH 9. Correns' curves, reproduced in Figure 1, are compiled as functions of pH and indicate that solubility equilibria of SiO_2 and $CaCO_3$ vary inversely with changes in pH, increasing pH favoring solution of SiO_2 and precipitation of $CaCO_3$, and decreasing pH favoring solution

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PLATE 2.—CARBONATE REPLACEMENT OF CRYSTALLINE SILICATE MINERALS

FIGURE 1.—Detrital feldspar cleavage fragment (white), extensively replaced by calcite (shades of gray). Dakota formation (Cretaceous) near Loveland, Colorado. Crossed nicols, 75×. The irregular patches of unreplaced feldspar are optically continuous and obviously are part of a once-continuous grain. The original outline of the grain has been destroyed by replacement. Collected by C. E. Breed.

FIGURE 2.—Intergrowth of quartz (white) and microcline (dark-gray rectangle) showing selective replacement of microcline by calcite (light gray). The quartz in this grain has been essentially unaffected by replacement. Minturn formation (Pennsylvanian), McCoy area, Eagle County, Colorado. Plain light, 40×.

FIGURES 3, 4.—Recrystallized oölite containing partially replaced silicate nuclei. Minturn formation (Pennsylvanian), Eagle County, Colorado. Figure 3 shows quartz nucleus (Q) deeply replaced by calcite (C) at lower left. Recrystallization has destroyed textural details of the oölites, but several "ghosts" of original oölites are recognizable. Plain light, 50×. Figure 4 shows an etched polished surface of specimen shown in Figure 3. Scattered silicate grains and silicate nuclei of oölites reveal deep etching owing to replacement. Authigenic silica is present as partially replaced oölite shells (white rings) and as small particles disseminated through interstitial calcite (small white flecks); 17×.

FIGURES 5, 6.—Replacement of microcrystalline quartz by dolomite along a fracture in a chert nodule. Montoya formation (Ordovician), Alamogordo, New Mexico. Figure 5 shows the fracture filled with clear dolomite (light gray) in optical continuity with dolomite euhedra that have replaced chert along the fracture walls. The dolomite of replacement origin is clouded by dusty inclusions derived from the replaced chert. Dolomite metacrysts are visible in the chert at some distance from the fracture (arrows). Plain light, 50×. Figure 6, same as Figure 5. Crossed nicols.

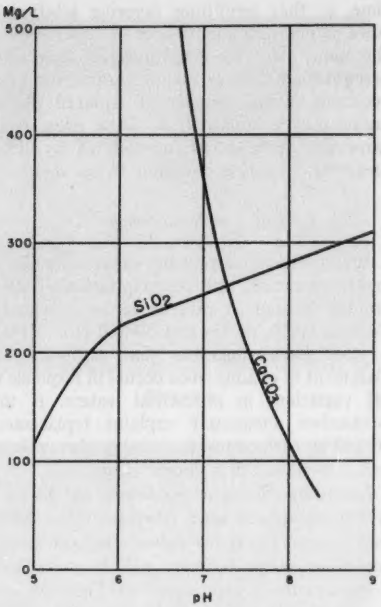

FIGURE 1.—SOLUBILITY OF AMORPHOUS SILICA AND CALCIUM CARBONATE IN WATERS OF VARYING pH
(From Correns, 1950)

of $CaCO_3$ and precipitation of SiO_2. On the basis of these data Correns suggests that interstitial waters saturated with $CaCO_3$ and SiO_2 at a particular pH would tend to precipitate $CaCO_3$ and dissolve more SiO_2 upon migration into an environment of higher pH, such as an area in which NH_3 is concentrated owing to bacterial activity. Under such conditions carbonate replacement of opal would occur. If, on the other hand, saturated waters migrate into an environment of lower pH, such as an area where CO_2 is concentrated owing to decay of organic matter, the replacement relationships would be reversed, and silica replacement of carbonates would occur.

This explanation of the cause of observed carbonate-silica replacement relationships seems reasonable in light of the silica solubility data compiled by Correns, but these data are not in agreement with more recent results obtained by other investigators (Alexander, Heston, and Iler, 1954; Krauskopf, 1956; Okamoto, Okura, and Goto, 1957). Laboratory studies by these writers indicate that solubility of amorphous silica increases rapidly with increase in pH ranges above pH 9, but in the geologically more common ranges below 9, solubility is nearly constant (Fig. 2). These data therefore indicate that Correns' explanation is inadequate unless alkalinity of for-

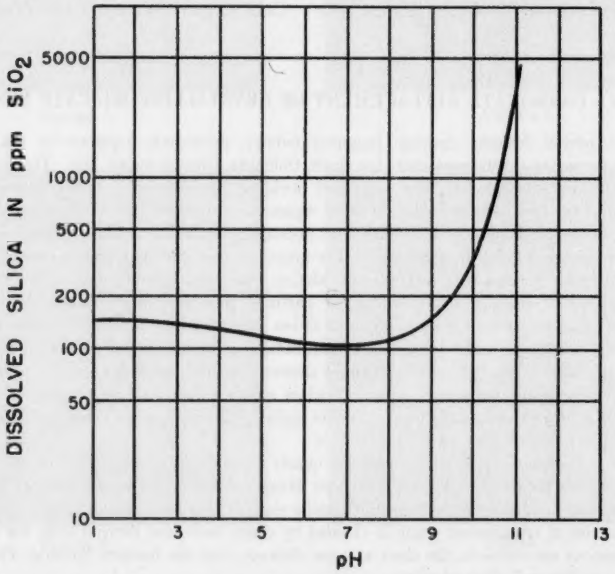

FIGURE 2.—SOLUBILITY OF AMORPHOUS SILICA IN WATERS OF VARYING pH
(From Alexander, Heston, and Iler, 1954)

mation waters at least locally rises above pH 9. Data available at present, however, indicate that pH values above 9 are not common in nature, and thus it appears that carbonate replacement of silicate minerals is caused by factors other than pH variations alone.

It is not presently known what other factors are involved or what role is played by each. Time may play an important role. High temperature and pressure resulting from deep burial possibly promote replacement, but apparently these factors are not essential in every occurrence, because extensively replaced silicate grains occur in sedimentary rocks that never have been deeply buried. Additional investigations are needed before the cause of these replacement phenomena will be adequately understood.

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TRANSCURRENT FAULTING AND VOLCANISM IN OWENS VALLEY, CALIFORNIA

BY L. C. PAKISER

ABSTRACT

In the Owens Valley region of California, volcanic activity of Cenozoic age was confined mainly to three areas near the ends of important faults. The volcanic eruptions seemingly took place in regions of relative tension, if the horizontal movement along these faults was left lateral. The deep depression of Owens Valley may have resulted from compression associated with left-lateral horizontal fault movement. The transfer of molten rock from beneath this deep depression laterally into the regions of tension and thence to the surface seems to account for the relief of abnormal stresses and the volume of the volcanic rocks.

CONTENTS

TEXT

	Page
Introduction	153
Acknowledgments	155
Regional geology	155
Theoretical considerations	155
Analysis of Owens Valley faults	157
Conclusion	158
References cited	159

ILLUSTRATIONS

Figure	Page
1. Map showing fault pattern and volcanic rocks of the Owens Valley region, California	154
2. Diagram illustrating how regions of tension and compression develop near the ends of transcurrent faults	156

INTRODUCTION

In the Owens Valley region of California, which is here extended to include Long Valley and Mono Basin (Fig. 1), volcanic activity of Cenozoic age was largely confined to three areas. These areas of volcanic activity, or volcanic fields, are found near the ends of important faults along which major vertical movement has taken place repeatedly through late Tertiary (?) to Recent times. In the Owens Valley earthquake of 1872, horizontal movement also took place along certain of the faults (Whitney, 1872; Hobbs, 1910; Freeman, 1932, p. 227–230; Hill, 1954; Gianella, 1957). However, these reports do not agree as to whether the horizontal movement was right lateral or left lateral. Richter (1958, p. 499–503) wrote a brief critical review of the evidence for horizontal fault movement in the 1872 earthquake. The purpose of this short paper is not to review these conflicting reports, nor to attempt to decide which observations were accurate and which were not, but the author leans toward the minority view that the horizontal movement was left lateral. The purpose of this paper is to review the association of the major faults with the volcanic fields to deduce a possible relation between the direction of horizontal fault movement and volcanic activity.

Mayo (1941) observed that fault movement between the granitic rocks of the Sierra Nevada and the sedimentary rocks farther east

> "... should open many channels for the extrusion of lava. It is therefore no surprise to find that volcanoes do occur along the Sierra Nevada front, but these eruptions are not evenly distributed along the base of the mountains... It is obvious that the volcanoes are clustered in certain favored areas."

Mayo (1941) also described a Recent mudflow welt west of Independence that has been offset by a fault about 8 feet in a left-lateral direction, although he inferred right-lateral motion of several hundred feet on the basis of a change in the course of a stream at the same place. The author has deduced that the favored areas mentioned by Mayo (1941) seem to be near the ends of transcurrent faults along which left-lateral horizontal movement probably has

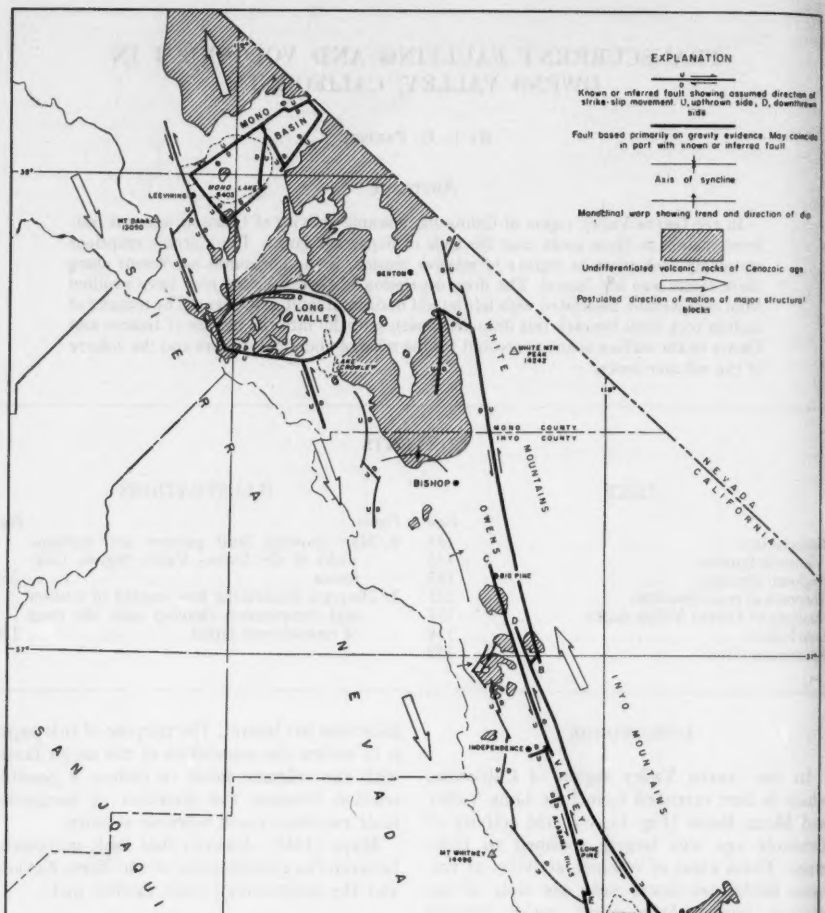

FIGURE 1.—MAP SHOWING FAULT PATTERN AND VOLCANIC ROCKS OF THE OWENS VALLEY REGION, CALIFORNIA

taken place. As nearly as possible, this deduction is based on observed relations between volcanic fields and faults, but the key to the puzzle—a demonstrated predominance of left-lateral movement—is missing. Therefore, the comments that follow are largely speculations. They are based almost wholly on kinematic concepts of strain and transfer of fluid mate-

rials; basic problems of available energy sources and stresses are left to future consideration.

Acknowledgments

The author is indebted to Vincent P. Gianella of the University of Nevada for much stimulating correspondence on the nature of faulting in the Owens Valley earthquake of 1872, and to James G. Moore of the U. S. Geological Survey for thought-provoking discussions with him of a similar problem of his involving failure in shear and quartz-vein injection associated with demonstrated left-lateral horizontal fault movement. The author wishes also to express his appreciation for the geologic information generously made available to him by Moore and by Paul C. Bateman of the U. S. Geological Survey. Although the author has benefited richly from the assistance of these geologists he alone is responsible for the principles set forth in this paper.

Regional Geology

Location of the faults and volcanic fields of the Owens Valley region (Fig. 1) were compiled from a number of sources, including in particular Knopf (1918), Gilbert (1941), Bateman and Merriam (1954), and maps published by the State of California Division of Mines. The regional relations and horizontal extent of these faults are based, in large part, on geophysical studies in Owens Valley, Long Valley, and Mono Basin (Pakiser and Kane, 1956; Pakiser, Press, Kane, and Warrick, 1958). Some details of the deformation along the Sierra Nevada front between Bishop and Independence were obtained from the unpublished geologic maps of Paul C. Bateman and James G. Moore of the U. S. Geological Survey. Some of the faults are inferred from physiographic evidence alone. The writer considers the tectonic pattern of the Owens Valley region shown in Figure 1 to be reliable on the whole and essentially complete on a regional scale.

The gravity data show (Pakiser and Kane, 1956) that Owens Valley subsided as a graben to receive an accumulation of about 8000 feet of predominantly clastic sediments of Cenozoic age. These clastic sediments are in general in high-angle fault contact with pre-Tertiary rocks throughout most, if not all, of their vertical extent. In southern Owens Valley between Independence and Owens Lake, the deepest wedge of Owens Valley is bounded on the east by a major fault marked by the escarpment of the Inyo Mountains and on the west by a large fault that includes the earthquake fault of 1872 along the eastern escarpment of Alabama Hills (Fig. 1). In northern Owens Valley from Bishop to Big Pine, the Owens Valley structure is bounded on the east by a major fault that lies along the escarpment of the White Mountains; on the west, between Bishop and Big Pine, both detailed geologic mapping (Paul C. Bateman, oral communication) and the gravity data show that the Sierra Nevada front is downwarped and not faulted.

Between Big Pine and Independence, the fault pattern of Owens Valley is marked by a series of overlapping faults, the most important of which terminate in this area at points B, C, and D (Fig. 1). The gravity data and areal geology show that pre-Tertiary rocks generally lie at relatively shallow depths in this area of intense volcanic activity. South of Owens Lake, the western bounding fault of the deepest wedge of Owens Valley lies along the Sierra Nevada front, but the eastern bounding fault diverges from the Inyo Mountains in this area.

The Owens Valley structure continues without interruption for about 25 miles north of Bishop.

The most striking features of the Owens Valley fault system are the near linearity of the fault along the front of the White and Inyo Mountains and the contrast between the simplicity of this fault and the irregularity of the system of faults and warps along the Sierra Nevada front.

The part of Mono Basin outlined by the faults that coincide roughly with the shore line of Mono Lake (Fig. 1) subsided about 18,000 feet to receive a deep accumulation of clastic sediments and volcanic materials of Cenozoic age (Pakiser, Press, Kane, and Warrick, 1958). Mono Basin has been interpreted as a volcanotectonic depression (or great caldera) that subsided along near-vertical faults following extrusion of volcanic materials from a magma chamber at depth (Pakiser, Press, Kane, and Warrick, 1958). Long Valley is a structure of the same type, as revealed by recent gravity data (Pakiser and Kane, 1956). These great structures are found in a major offset of the Sierra Nevada front.

The Cenozoic volcanic activity of the Owens Valley region was prolonged and complex. Excellent detailed accounts of the volcanic history have been written by Knopf (1918), Gilbert (1938; 1941), and Mayo (1941).

Theoretical Considerations

All transcurrent faults must terminate somewhere. Rocks may be distorted or strained on

either side near the ends of a transcurrent fault along which some finite horizontal movement has taken place (Fig. 2A). The fault displacement at both ends must be zero, and at some unspecified distance from each of these ends the displacement of one side relative to the other must increase to the full amount of horizontal movement w. On one side near each end (the side on which the displacement arrow points away from the end) the rocks will be extended, and a region of relative tension will come into existence. On the opposite side near each end (the side on which the displacement arrow points toward the end) the rocks will be effectively shortened, and a region of compression will result. These regions will be disposed alternately, so that a region of relative tension and a region of compression will exist on each side of the fault.

Failure in tensile fracture (at shallow depths) and volcanic eruption will be favored in the regions of relative tension. Failure in shear (either as secondary strike-slip—or splay—faults or thrust faults), or by compressive folding, will be favored in the regions of compression.

If two parallel transcurrent faults with the same direction of displacement are arranged en echelon, and if they extend outward in opposite directions, either a region of relative tension or a region of compression will be developed between them, depending on whether the region between them tends to be extended or shortened (Fig. 2, B and C). The two faults may, of course, overlap. Failure in tension or compression will be similar to that for a single fault. Where the region between the faults is one of compression, splay faults (Anderson, 1951, p. 166–167), thrust faults, or faults of the Brenta type (De Sitter, 1956, p. 167–169) may form. Regions of relative tension again will presumably favor volcanic eruption. It may readily be shown by a simple experiment using a card with two cuts representing overlapping faults that a small horizontal displacement tending to shorten the strip between the faults will cause a large vertical displacement of the strip.

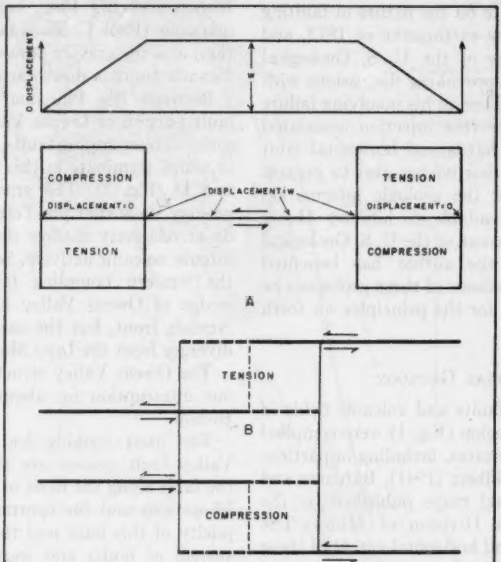

FIGURE 2.—DIAGRAM ILLUSTRATING HOW REGIONS OF TENSION AND COMPRESSION DEVELOP NEAR THE ENDS OF TRANSCURRENT FAULTS

In the discussion above, the term tension has been qualified only by the modifier relative. Except relatively near the surface, absolute tension may not exist because the lithostatic pressure determined by the weight of the rocks must exceed the relative tensile stress at great (and here unspecified) depths, and the rocks will be everywhere in compression. Nevertheless, the compressive stresses parallel to a fault may be greatly reduced, and the rocks may fail in tensile fracture as a result of the upward propulsion of magma in the manner described by Anderson (1951, p. 22–28). The quantitative aspects of rock deformation near the ends of a transcurrent fault have not been considered here because the purpose has been only to identify the regions of relative tension and compression.

Analysis of Owens Valley Faults

These principles may be applied to the volcanic fields in the Owens Valley region (Fig. 1). In the volcanic field near the southern end of the Inyo Mountains just north of point A (Knopf, 1918) the volcanic eruptions took place on the east side of the bounding fault. A region of relative tension presumably existed in the area of these volcanic eruptions, and it may be inferred, therefore, that the Inyo Mountains moved north relative to Owens Valley and that the bounding fault is a left-lateral transcurrent fault.

The three vents for the small volcanic field that flowed out of the Inyo Mountains between Big Pine and Independence (Fig. 1) are located on the east side of the easternmost bounding fault (Mayo, 1941), and the bounding fault system consists of two offset segments in this area. The volcanic eruptions took place near and east of the end of the fault (point B) that continues northward without interruption nearly to the northern limit of the White Mountains. Assuming again that the volcanic eruptions took place in a region of relative tension, the writer infers that the White Mountains and Inyo Mountains moved north relative to Owens Valley and that the bounding fault is a left-lateral transcurrent fault.

The bounding fault or fault system of the White Mountains and Inyo Mountains is the most persistent feature in the Owens Valley region, and it strikes in the same direction as the trend of Owens Valley as a whole. Therefore the entire Owens Valley system of faults may be a great left-lateral shear zone.

Examination of the outcrops of volcanic rock south of Big Pine and north of Independence shows that the volcanic area just south of Big Pine (Fig. 1) is near the end of a fault (point C) that passes northward into the warp mapped by Bateman and confirmed by the gravity evidence. A region of relative tension presumably existed on the west side of this fault near its northern end, and it is again reasonable to infer left-lateral movement on this fault, which is an extension of the fault east of Independence mapped by Knopf (1918). This fault is parallel to and may join the earthquake fault of 1872 that bounds Alabama Hills near Lone Pine, and therefore left-lateral movement may be inferred for that fault also. The irregular system of faults along the Sierra Nevada escarpment in this area may have provided channels along which lava could have been propelled to the surface. All the vents for the Big Pine volcanic field are located on the west side of the fault that terminates at point C (Mayo, 1941).

Thus the volcanic activity and the fault pattern in the area between Big Pine and Independence seem strongly to suggest left-lateral movement. The writer suggests that the Sierra Nevada moved south with respect to Owens Valley.

The strip between the faults terminating at points A and C (Fig. 1) should be a region of compression if the movement along both faults was left lateral, for this strip would tend to be shortened, *i.e.*, points A and C would tend to be brought closer together. (*cf.* Fig. 2C.) This would also be true of the shorter segments of this strip between the faults terminating at points B and C and B and D (Fig. 1). The shortening or compression of this region of overlapping faults may explain the fact that it was in part of this region of overlap, between Independence and Owens Lake, that the pre-Tertiary floor of the deepest wedge of Owens Valley was depressed far below sea level. Possibly some of the short faults that project into Owens Valley in this strip are reverse faults. If a source of molten basalt existed beneath this downward-pressing strip, presumably the basalt would migrate laterally, contemporaneously with the faulting, into the adjacent regions of relative tension and thence to the surface. In effect, the magma would be pumped to the surface. Thus the downward pressure beneath this deep wedge would be relieved by lateral migration of basalt, the regions of relative tension would tend to be returned to a normal stress condition by in-

truded basalt, and the volume of the volcanic rocks would be adequately accounted for by balancing them and the intruded basalt of unknown volume against the volume of the depressed wedge.

Another possible interpretation of this deduction is that Alabama Hills may have been elevated, either by folding or reverse faulting, as a result of compressive stresses acting parallel to the earthquake fault of 1872 in the wedge north and west of point E. This would seemingly be a region of compression (Fig. 2A).

The broader area containing Mono Basin and Long Valley can be compared with the similar idealized example (Fig. 2B). This area, in which the Sierra Nevada front has been so prominently offset, seems to have been a region of relative tension if the Sierra Nevada moved south with respect to the area to the east, as indicated by the arrows showing the postulated direction of motion of major structural blocks (Fig. 1). As these arrows suggest, this area may have had a tendency to be stretched or pulled apart. It was also an area of prolonged, intense, and complex volcanic activity. From this area the Bishop tuff (Gilbert, 1938) was extruded, the Mono Craters erupted explosively (Williams, 1932; Gilbert, 1941; Putnam, 1949), and a sequence of volcanic rocks, ranging in age from early Pliocene(?) to late Pleistocene and in composition from basalt to rhyolite, was emplaced. Two unusual structures in this region, the volcanotectonic depressions of Mono Basin and Long Valley, subsided to depths far below sea level and were filled to the brim with clastic sediments and volcanic ejecta (Pakiser, Press, Kane, and Warrick, 1958).

The writer infers that high internal pressure in an intracrustal magma chamber (perhaps exerted mainly by dissolved water) propelled the silicic volcanic rocks of Long Valley and Mono Basin to the surface as this area was relatively stretched. The basaltic magmas probably came from a deeper source; they may have been "pumped" to the surface from below adjacent regions of compression, or they may have risen in part by hydrostatic pressure. Possibly magma was generated at depth in regions of tension as a result of relief of lithostatic pressure and the inward migration of water and other volatiles into such regions (Uffen, 1959).

The writer assumes that the Sierra Nevada block has been moving south. The major Arvin-Tehachapi earthquake of 1952 took place along the White Wolf fault in and west of the Tehachapi Mountains south of the Sierra and north of and parallel to the Garlock fault (Buwalda and St. Amand, 1952; Oakeshott, 1955). Benioff (*in* Oakeshott, 1955, p. 199–202) has interpreted the failure of the White Wolf fault as a result of a combination of high-angle reverse faulting and left-lateral strike-slip movement, a clear indication of failure as a result of compression, as would be required by a south-moving Sierra block. The northern Sierra was, in Cenozoic time (and may still be), a region of considerable volcanic activity, whereas the southern Sierra is virtually devoid of volcanic rocks of Cenozoic age. This uneven distribution of volcanic rocks in the Sierra provides further evidence that the Sierra block has been moving south. The volcanic field in the embayment of the Sierra Nevada front near Blairsden seems to agree with the principle deduced in this paper, and in this area Durrell (1950) mapped a fault with more than 3 miles of left-lateral strike-slip displacement. When this fault as described by Durrell (Written communication) was plotted on a geologic map of California, it was found to strike within a few degrees of the trend of Owens Valley.

The uplift of the Sierra Nevada took place only in minor part along the faults that bound the deepest wedge of Owens Valley. This uplift was probably caused by vertical forces resulting from isostatic readjustment, as has been proposed by Oliver (1956). The Owens Valley shear zone probably existed before the late Pliocene and early Pleistocene uplift of the Sierra, and Owens Valley may have begun to subside and receive sediments at some earlier time.

Conclusion

The author believes it highly probable that Owens Valley is a great left-lateral shear zone, as was suggested recently on other grounds by Moody and Hill (1956). Vincent P. Gianella (Written communication) has advised the author that the horizontal movement in the earthquake of 1872 along the major fault east of Independence was movement of "undoubted left-lateral" direction on the basis of his recent examination of the fault and his review of the published evidence. (Since this paper was written, Gianella has called to the author's attention two other places south of Lone Pine where there is evidence of left-lateral fault

movement.) The volcanic activity in the Owens Valley region seems to be controlled in large part by perhaps rather small left-lateral horizontal movements along faults that extend essentially through the earth's crust. The Dead Sea rift zone (Quennell, 1958) exhibits a relation to volcanism seemingly similar to that in Owens Valley.

The author wishes to emphasize again the speculative nature of this short paper and to acknowledge that future observations, in the Owens Valley area and elsewhere, may not bear out these conclusions.

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LATE PLEISTOCENE GLACIATION OF EASTERN NEWFOUNDLAND

BY STUART E. JENNESS

ABSTRACT

Melting of the late Pleistocene ice sheet has left a semicircle of glacial deposits around the coast of eastern Newfoundland west of the Avalon Peninsula. A discontinuous end moraine accentuates the curvature and separates ground moraine on the coastal side (outer drift zone) from slightly younger eskers, kames, and ground moraine on the inner side (inner drift zone). The distribution of ground moraine, glacial striations, indicator boulders, and glaciated bedrock ridges in the two drift zones indicates that an ice sheet once covered all eastern Newfoundland west of the Avalon Peninsula. The ice center was somewhere in western Newfoundland. At about the same time local ice caps existed on the Avalon Peninsula. Small valley glaciers persisted in a few elevated regions of the north and south coasts west of longitude 55° after the retreat of the main ice sheet.

Eastern Newfoundland appears to be tilting upward toward the northwest, with a zero isobase passing through Trinity and Placentia bays. Northwest of this axis the coast is emerging; southeast of it (on the Avalon Peninsula) the coast is submerging. The zero isobase veers west along the south shore of Newfoundland, parts of which are submerging, parts emerging.

CONTENTS

TEXT

	Page
Introduction	161
Glaciation of eastern Newfoundland west of the Avalon Peninsula	163
General statement	163
Glacial dating	163
Source of the ice	163
Direction of ice movement	164
Zonal distribution of glacial deposits	166
Outer drift zone	166
Physiographic features	166
Glacial deposits	168
Inner drift zone	171
Physiographic features	171
Glacial deposits	173
Glaciation on the Avalon Peninsula	174
Postglacial coastal movements	175
Summary and conclusions	177
References cited	178
Sources of reference for Figure 2	178

ILLUSTRATIONS

Figure	Page
1. Map of eastern Newfoundland showing locations of places mentioned in text	162
2. Map of eastern Newfoundland showing orientation of glacial striations and positions of eskers	164
3. Map of eastern Newfoundland showing distribution of late Pleistocene glacial features	165
4. Cross-section sketch of Gander Lake	167
5. Depth contour map of Trinity Bay showing the steepness of part of its east side and great depth of the bay	169
6. Map showing location of elevated alluvial terraces in eastern Newfoundland	172
7. Ice centers on the Avalon Peninsula	175
8. Isobases in eastern Newfoundland, based upon upper levels of outwash deltas	176

TABLE

Table	Page
1. Locations of outwash trains and deltas in eastern Newfoundland	170

INTRODUCTION

Most of our present knowledge of the glacial geology of eastern Newfoundland stems from the early works of Daly (1921), Coleman (1926), and MacClintock and Twenhofel (1940). Their observations were restricted almost entirely to points around the coast and along the railway, as the interior was relatively inaccessible. Much of the interior has since become accessible through the construction of roads and the introduction of commercial

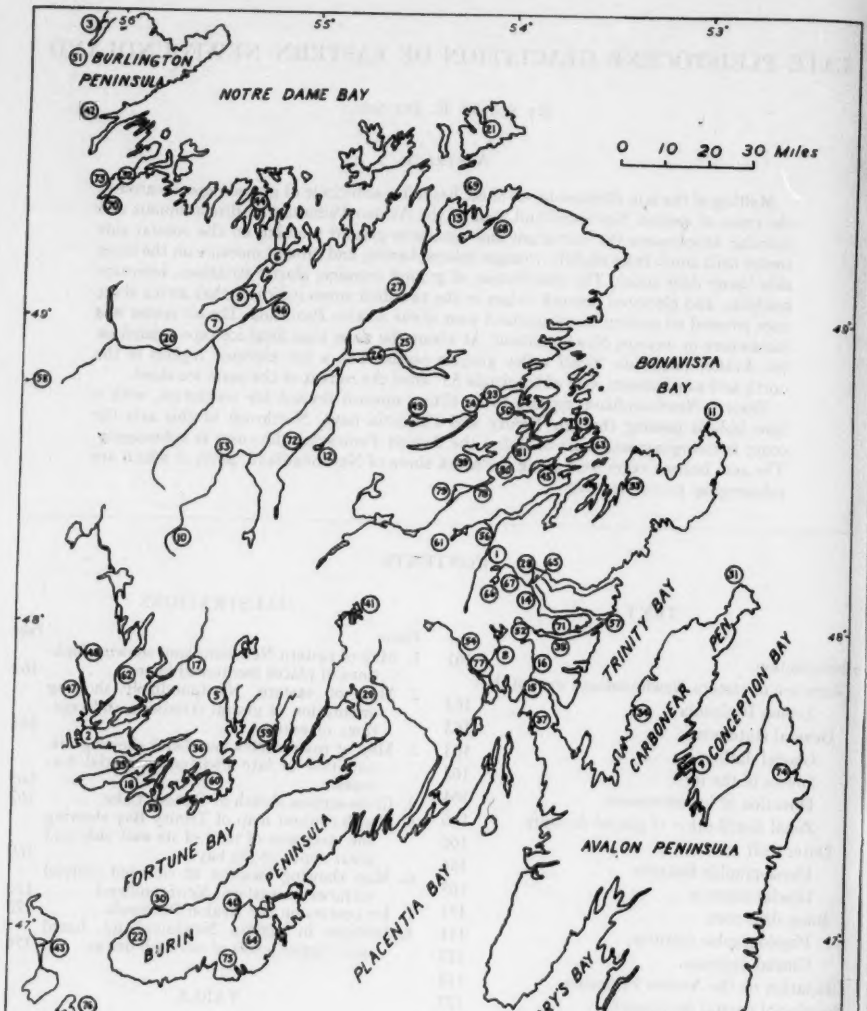

FIGURE 1.—MAP OF EASTERN NEWFOUNDLAND SHOWING LOCATIONS OF PLACES MENTIONED IN THE TEXT

1. Andrews Pond
2. Baie (Bay) d'Espoir
3. Baie Verte
4. Bay de Grave
5. Bay du Nord River
6. Bay of Exploits
7. Bishops Falls
8. Black River
9. Botwood
10. Burnt Hill
11. Cape Bonavista
12. Caribou Pond
13. Carmanville
14. Clarenville
15. Come by Chance
16. Come by Chance River
17. Conne River
18. Dawson's Cove
19. Eastport
20. Exploits River
21. Fogo Island
22. Fortune Valley
23. Freshwater Bay
24. Gambo

(*Continued at foot of next page*)

float aircraft. In addition, complete aerial photograph coverage and complete topographic map coverage are now available for eastern Newfoundland. Yet, despite these new mapping aids, it appears from two recent publications (Prest, 1957; Wilson *et al.*, 1958) that little glacial information has been added to the published literature since 1940.

The purpose of this paper is to describe the distribution and types of glacial features in eastern Newfoundland, to interpret the features described, and to stimulate renewed interest in the glacial geology of Newfoundland. The glacial data presented herein have been obtained from publications of the Geological Survey of Newfoundland and the Geological Survey of Canada, from various geological journals, from several unpublished manuscripts and maps, and from the personal observations of the writer during mapping of bedrock lithologies and structures in eastern Newfoundland over the course of five field seasons.

"Eastern Newfoundland" as used in this paper refers to that part of the island east of longitude 56° and south of latitude 50°. All places mentioned are shown on Figure 1.

The writer expresses his appreciation to Prof. R. F. Flint of Yale University and to Dr. V. K. Prest of the Geological Survey of Canada for their helpful suggestions during the preparation of the manuscript. The writer alone is responsible for the interpretations given.

Glaciation of Eastern Newfoundland West of the Avalon Peninsula

General Statement

Glacial dating.—All eastern Newfoundland was glaciated during Pleistocene times. In the absence of evidence to the contrary the glaciation has been assumed to have occurred in late Pleistocene (Wisconsin) times (MacClintock and Twenhofel, 1940). This generalized date has not been modified in the past two decades, but a more accurate date for the time of glaciation could now be obtained by use of radioactive carbon-14 methods. Ample material for carbon-14 age determinations is available in the many bogs and shallow ponds throughout Newfoundland. This material, if carefully selected, would also yield pollen for studies that would provide insight into climatic conditions since deglaciation. Such studies have been made in Nova Scotia by Livingstone and Livingstone (1958). It is hoped that studies of this type will soon be undertaken in Newfoundland.

Source of the ice.—Although the source(s) of the late Pleistocene ice sheet cannot yet be accurately located, the directions of striations, composition of till boulders, and the general distribution of glacial deposits indicate that the ice moved east from the western half of the island. Murray (1955, p. 272) suggests that the Annieopsquatch Mountains near Victoria Lake constituted one ice center, but whether this center fed eastern Newfoundland is not known.

---

25. Gander
26. Gander Lake
27. Gander River
28. Georges Brook
29. Gisburn Lake
30. Grand Bank
31. Grates Point
32. Hall's Bay
33. Harbour Breton
34. Heart's Delight
35. Hermitage
36. Hermitage Peninsula
37. Isthmus of Avalon
38. Long Beach
39. Maccles Pond
40. Main Brook
41. Meta Pond
42. Middle Arm of Green Bay
43. Miquelon
44. New Bay
45. Newman Sound
46. Norris Arm
47. North Bay
48. North Bay Brook
49. North Pond
50. Northwest Arm Alexander Bay
51. Northwest Brook of Baie Verte
52. Northwest Brook of Trinity Bay
53. Northwest Gander River
54. Pipers Hole River
55. Plate Cove
56. Port Blandford
57. Random Sound
58. Red Indian Lake
59. Rencontre Lake
60. Salmonier Cove Pond
61. Salmon Pond
62. Salmon River in Baie d'Espoir
63. Sandy Cove
64. Shearstick Brook
65. Shoal Harbour
66. Shoal Harbour Pond
67. Shoal Harbour River
68. Shoal Pond
69. Sir Charles Hamilton Sound
70. South Brook
71. Southwest Arm Random Sound
72. Southwest Gander River
73. Springdale
74. St. John's
75. St. Lawrence
76. St. Pierre
77. Swift Current
78. Terra Nova
79. Terra Nova Lake
80. Terra Nova River
81. Traytown

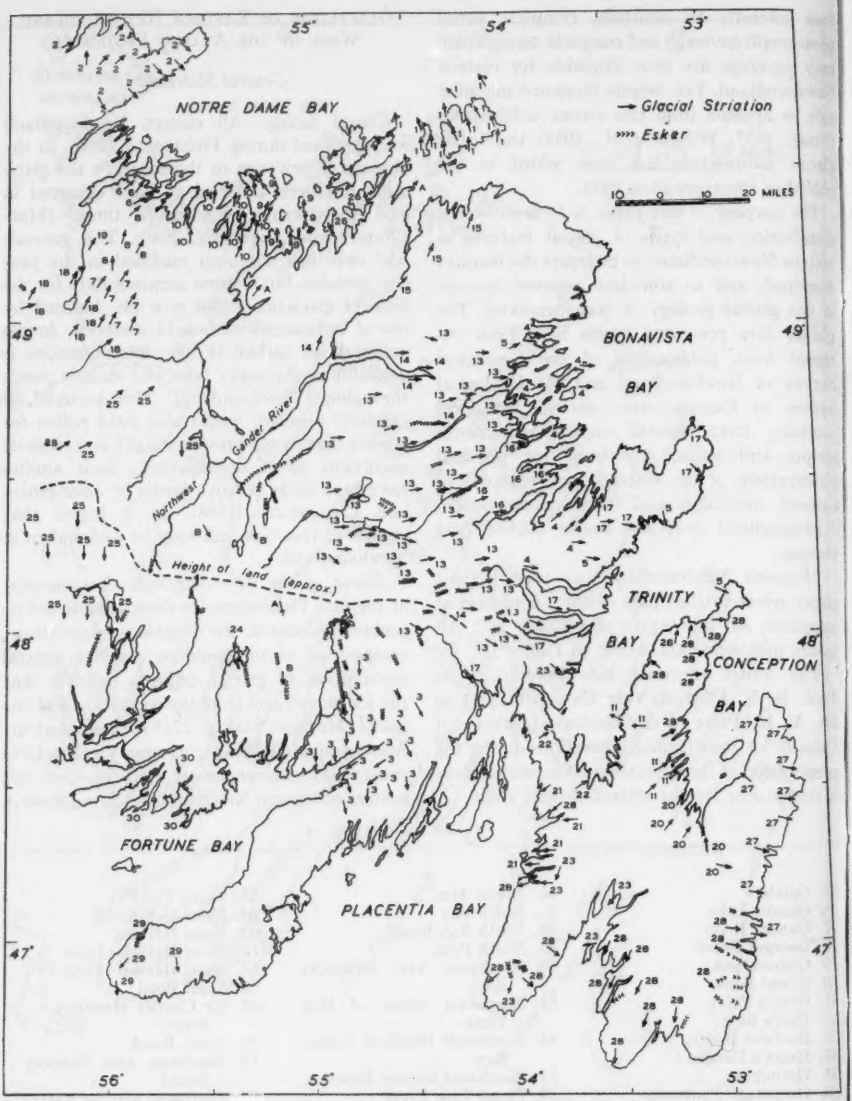

FIGURE 2.—MAP OF EASTERN NEWFOUNDLAND SHOWING ORIENTATION OF GLACIAL STRIATIONS AND POSITIONS OF ESKERS

Numbers denote references in list following References Cited. "A" denotes striations taken from MacClintock and Twenhofel (1940); "B" denotes striations taken from three unpublished maps of the Department of Mines and Resources, St. John's, Newfoundland.

Direction of ice movement.—The ice in eastern Newfoundland moved generally coastward, as indicated by the orientation of striations, positions of indicator boulders, and shapes of glaciated ridges. In a few places topographic lineaments, notably at the head of Fortune Bay and along the south side of Bonavista Bay, modified the direction of flow.

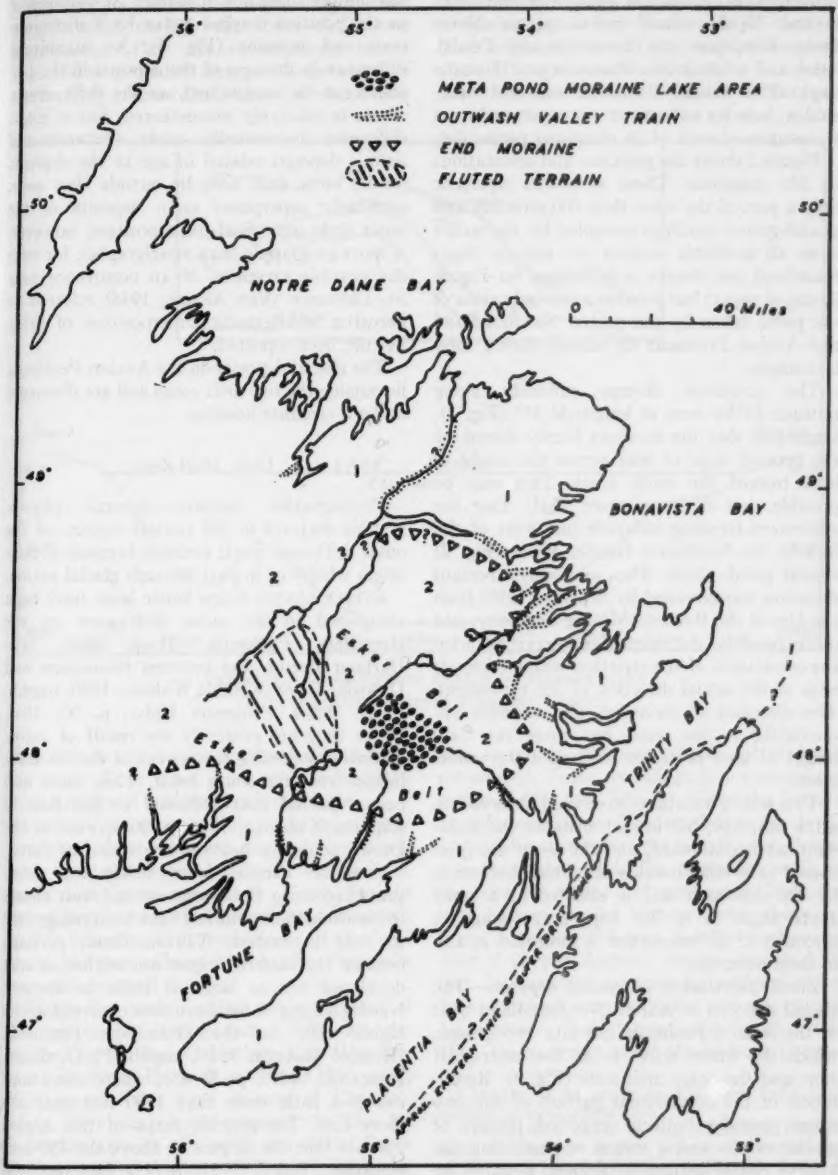

FIGURE 3.—MAP OF EASTERN NEWFOUNDLAND SHOWING DISTRIBUTION OF LATE PLEISTOCENE GLACIAL FEATURES
End moraine separates inner drift zone (2) from outer drift zone (1)

Movement in general was radially outward from central Newfoundland—north across Notre Dame Bay, east into Bonavista and Trinity bays, and south across Placentia and Fortune bays. The Avalon Peninsula was not overridden; here ice movement was coastward from the centers of each of its elongated peninsulas.

Figure 2 shows the positions and orientations of 236 striations. These striations represent only a part of the more than 600 striation and glacial-groove readings compiled by the writer from all available sources for eastern Newfoundland (*see* Source of references for Figure 2, end of paper) but provide a fair indication of the paths taken by the central Newfoundland and Avalon Peninsula ice masses during their last stages.

The striations diverge curiously along latitude 48°30′ west of longitude 55° (Fig. 2), suggesting that the ice sheet locally flowed *up* the present slope of land across the height of land toward the south coast. This may be possible, but it seems more likely that the southward-trending striation just west of the fork in the Northwest Gander River (Fig. 2) should point north. This seemingly aberrant direction was obtained by Murray (1955) from the files of the Buchans Mining Company, and quite possibly the original observer recorded the orientation of the striation without knowledge of the actual direction of ice movement. The direction of striations also changes immediately to the west, but there the local height of land probably influenced the movement.

Two sets of striations have been observed on a few outcrops, but in each instance the dominant set is coastward, and the lesser set (presumed to be older because of partial obscurance by the dominant set) is oriented at a small acute angle to it. No important change in direction of ice movement is indicated in any of these outcrops.

Zonal distribution of glacial deposits.—The glacial deposits in eastern Newfoundland west of the Avalon Peninsula fall into two groups, which the writer refers to as the inner drift zone and the outer drift zone (Fig. 3). Recognition of the semicircular pattern of the two zones provides both a simplified picture of glacial events and a means of predicting the kind of glacial features to be found in adjoining unstudied areas.

Most of the glacial deposits in the inner drift zone formed when the ice front had retreated from its maximum position just outside the coast line of eastern Newfoundland to the position marked today by a discontinuous end moraine (Fig. 3). No significant difference in the ages of the deposits in the two zones can be ascertained, as the drift everywhere is relatively unweathered, but a slight difference undoubtedly exists. Outwash and deltaic deposits related in age to the deposits of the inner drift zone lie outside that zone, seemingly superposed upon deposits of the outer drift zone. Such superposition, however, is more geographic than stratigraphic, for with the possible exception of an occurrence near St. Lawrence (Van Alstine, 1948) referred to hereafter stratigraphic superposition of drifts has not been reported.

The glacial deposits on the Avalon Peninsula lie outside the two drift zones and are discussed under a separate heading.

Outer Drift Zone

Physiographic features.—Several physiographic features in the coastal regions of the outer drift zone merit mention because of their origin wholly or in part through glacial action.

KETTLE LAKES: A few kettle lakes have been recognized in the outer drift zone on the Hermitage Peninsula. These occur near Harbour Breton and between Hermitage and Dawsons Cove (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 90, 101), where they are evidently the result of valley glaciation following the retreat of the ice sheet inland from the south coast. They have also been reported near Botwood in the Bay of Exploits (Coleman, 1926, p. 206), west of the known boundary between the two drift zones.

CIRQUES: Cirques were recognized many years ago along the southwest and west coasts of Newfoundland, but few have been recognized to date in eastern Newfoundland, perhaps because the eastern cirques are neither as well developed nor as large as those in western Newfoundland. A few have been observed in the higher hills on the Hermitage Peninsula (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 81–85), where elevations rise to a little more than 1000 feet near the shore line. Topographic maps of this region indicate that the cirques lie above the 750-foot elevation. Baird (1951, p. 11) has reported bowl-like depressions, which may be cirques, on the north coast of Newfoundland along the Burlington Peninsula.

The small size of the cirques and their

restriction to regions of high altitude and rugged relief outside the inner drift zone suggest that they are of relatively recent origin. Most probably they are the products of small valley glaciers that existed for some time after the retreat of the ice front inland from the coast. Even today snow and ice linger in these areas well into the summer months.

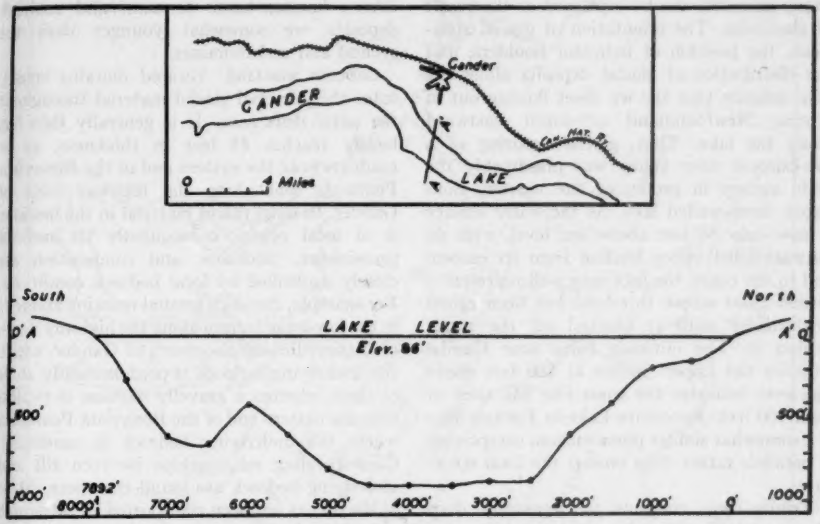

FIGURE 4.—CROSS-SECTION SKETCH OF GANDER LAKE
(Data from Department of Transport, Gander)

FJORDS AND FJARDS: Fjords are extensively developed along the south coast of Newfoundland from the Burin Peninsula westward. A particularly good example is North Bay in Bay d'Espoir, which has a depth near its mouth of 390 fathoms (2340 feet). Some of the bays along the east side of the Avalon Peninsula may also be small fjords.

Fjards (or fiards), which are re-entrants of the sea having low glaciated sides, are developed in the Bay of Exploits and New Bay areas in Notre Dame Bay (Heyl, 1936, p. 2; 1938, unpub. ms., Geol. Survey of Canada). They have also been reported in the St. Lawrence area on the south end of the Burin Peninsula (Van Alstine, 1948, p. 6). Many of the embayments around the east coast of Newfoundland between the St. Lawrence area and the Bay of Exploits may also be fjards.

DEEP WATER BASINS: Two unusually deep large bodies of water in eastern Newfoundland —Gander Lake and Trinity Bay—seem to have originated by or been modified by glaciation. The great depth of Gander Lake[1] has puzzled the writer since 1952, but he has been able to obtain only one reliable set of depth recordings[2]. These recordings give a maximum depth of 900 feet (Fig. 4). The lake bottom at this location is 815 feet below sea level. Local fishermen report the lake is even deeper near its western end. The containing walls of the lake are steep, and total relief, from the lake bottom to the highest point above its shore (elevation 695 feet, 6 miles west-southwest of Gander Airport) slightly exceeds 1500 feet. Its orientation across the regional structure, its steep, narrow walls, and the curious below-water "peak" near the south shore (Fig. 4), occurring where the bedrock is prac-

[1] Depths of as much as 5000 feet have been quoted by local inhabitants.

[2] These were made by the Canadian Department of Transport in 1955 prior to the installation of electronic aircraft-landing equipment south of one of the airport runways at Gander. Soundings were taken on a calm day by lowering a heavily weighted nylon cord from a small boat; the position was established for each station by several transits on shore.

tically flat-lying, are puzzling. Structure partly controls the major bend in the lake just west of Gander, as the sedimentary rocks there are locally flattened and intruded by a north-plunging lobe of granite, but it does not explain the pronounced topographic features. Most of these, however, can be explained as the result of glaciation. The orientation of glacial striations, the position of indicator boulders, and the distribution of glacial deposits along the lake indicate that the ice sheet flowing out of central Newfoundland advanced eastward along the lake. Thus, glacial scouring of a pre-existent river valley was presumably the main agency in producing the narrow, elongated, steep-walled lake. As the water surface is now only 85 feet above sea level, with an outwash-filled valley leading from its eastern end to the coast, the lake may well represent a former fjord whose threshold has been raised isostatically until it blocked off the water behind it. The outwash delta near Gambo with its flat upper surface at 100 feet above sea level indicates the coast line has risen at least 100 feet. Rencontre Lake in Fortune Bay is a somewhat similar phenomenon, except that it parallels rather than crosses the local structure.

Trinity Bay contains an unusually deep elongated basin (Fig. 5), which would appear to be the result of glacial deepening, as in the case of the fjordic bays on the south coast. However, ice movement appears to have been at right angles to the northeast orientation of the basin, and the shape of the bottom profile is not typical of a glaciated valley. The bay contains a straight, steep slope on its eastern side, and has a maximum recorded depth of 1920 feet east of the mouth of Random Sound (U.S. Navy Hydrographic Office Chart H.O. 1102). These features, combined with the orientation (N.30°E.) of the eastern slope, which parallels the orientation of major strike faults in the Conception Bay area to the east (Rose, 1952, p. 40) and in the Clarenville area to the west (Hayes, 1948, geologic map), seem to indicate a major submarine fault scarp. Structure rather than glaciation thus appears to have effected the development of the unusual bottom topography of the bay, although some modification by ice action probably did occur.

Glacial deposits.—Most of the glacial deposits in the outer drift zone are the products of advancing ice. They consist principally of relatively unstratified ground moraine, largely of local origin, but include a few small end moraines, drift ridges, eskers, kames, kame terraces, and large patches of glacial outwash trending seaward from the margin of the inner drift zone. Most of the outwash deposits terminate as elevated deltas at the coast. The eskers, kames, kame terraces, and outwash deposits are somewhat younger than the ground and end moraines.

GROUND MORAINE: Ground moraine constitutes the principal glacial material throughout the outer drift zone. It is generally thin but locally reaches 25 feet in thickness, as in roadcuts near the eastern end of the Bonavista Peninsula and along the highway west of Gander. In most places material in the moraine is of local origin; consequently its surficial prominence, thickness, and composition are closely controlled by local bedrock conditions. For example, clay-rich ground moraine abounds in the low-lying terrain along the highway north of Clarenville and also north of Gander, where the underlying bedrock is predominantly shale or slate, whereas a gravelly moraine is typical near the eastern end of the Bonavista Peninsula where the underlying bedrock is sandstone. Corresponding relationships between till and underlying bedrock are found elsewhere. Most of the till has not been transported far, although a few porphyritic granite boulders near Eastport, in Bonavista Bay, are at least 25 miles east of known outcrops. Such boulders are easily recognized, hence can be spotted for many miles from their source, but are by no means abundant for more than a mile or two beyond the margins of the granite outcrops.

Although the ground moraine in the outer drift zone is somewhat older than that in the inner drift zone, there is no recognizable physical characteristic indicating any important age difference. As no carbon-14 determinations have been made on material from anywhere in Newfoundland,[3] the only criteria of age available are (1) the pattern of distribution of the glacial deposits, and (2) the unweathered aspect of all till in eastern Newfoundland. No till older than the ground moraine in the outer drift zone has yet been recognized.

END MORAINE: Although most of the outer limit of the glacial deposits in the outer drift

[3] Basal peat collected by E. P. Henderson from a 17-foot-thick bog near St. John's has yielded a radiocarbon date of 7400 ± 150 years (*reported in* Olson and Broecker, 1959). This is interpreted as a minimum age for deglaciation in the St. John's area. It is the first radiocarbon age from Newfoundland.

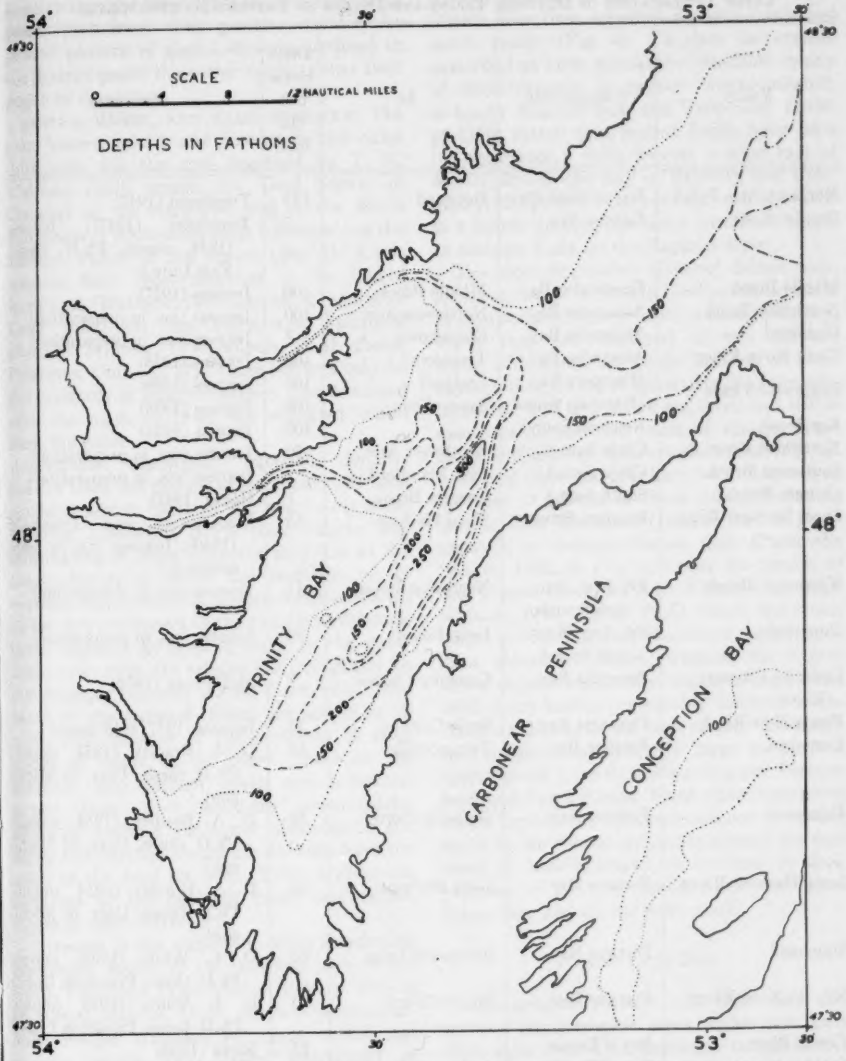

FIGURE 5.—DEPTH CONTOUR MAP OF TRINITY BAY SHOWING THE STEEPNESS OF PART OF ITS EAST SIDE AND GREAT DEPTH OF THE BAY
(Data from U.S.N. Hydrographic chart 1102)

zone lies in the ocean, making recognition of an end moraine impossible, a few local end moraines have been observed. Two small ones on Shoal Pond (near Carmanville in Sir Charles Hamilton Sound) (Jenness, 1954, unpub. Ph.D. thesis, Yale Univ., p. 140), probably developed when the ice front was retreating inland from the coastal area. Several others on the Hermitage Peninsula in Fortune Bay (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 90) probably represent the termini of local valley glaciers that existed in this part of

TABLE 1.—LOCATION OF OUTWASH TRAINS AND DELTAS IN EASTERN NEWFOUNDLAND

Valley	Empties into	At	Delta elevation (In feet)	Reference
Northern Arm Brook	Bay of Exploits	Botwood	175	Twenhofel (1947)
Gander River	Gander Bay	..	?	Twenhofel (1947); Jenness (1954, unpub. Ph.D. thesis, Yale Univ.)
Middle Brook	Freshwater Bay	Middle Brook	100	Jenness (1957)
Northwest Brook	Alexander Bay	Northwest Arm	100	Jenness (ms. in preparation)
Unnamed	Alexander Bay	Glovertown	?	Jenness (ms. in preparation)
Terra Nova River	Alexander Bay	Traytown	100	Jenness (1958)
Southwest Tickle	Eastport Bay	Eastport	100	Jenness (1958)
	Newman Sound	Sandy Cove	100	Jenness (1958)
Big Brook	Newman Sound	..	100	Jenness (1958)
Northwest River	Clode Sound	Northwest Arm	50	Jenness (ms. in preparation)
Southwest Brook	Clode Sound	Port Blandford	?	Jenness (ms. in preparation)
Georges Brook	Smith Sound	Georges Brook	?	Hayes (1948)
Shoal Harbour River	Random Sound	Shoal Harbour	55	MacClintock and Twenhofel (1940); Jenness (ms. in preparation)
Northwest Brook	SW Arm, Random Sound	Northwest Brook	45	Jenness (ms. in preparation)
Unnamed	SW Arm, Random Sound	Long Beach	50	Jenness (ms. in preparation)
Come by Chance River	Placentia Bay	Come by Chance	?	McCartney (1958)
Pipers Hole River	Placentia Bay	Swift Current	35	Jenness (1957, field notes)
Unnamed	Fortune Bay	Terrenceville	65	D. A. Bradley (1954, unpub. Ph.D. thesis, Univ. of Michigan)
Unnamed	Fortune Bay	Grand le Pierre	55	D. A. Bradley (1954, unpub. Ph.D. thesis, Univ. of Michigan)
Long Harbour River	Fortune Bay	Long Harbour	65	D. A. Bradley (1954, unpub. Ph.D. thesis, Univ. of Michigan)
Unnamed	Fortune Bay	Rencontre Lake	65	D. E. White (1939, unpub. Ph.D. thesis, Princeton Univ.)
Bay du Nord River	Fortune Bay	Bay du Nord	65	D. E. White (1939, unpub. Ph.D. thesis, Princeton Univ.)
Conne River	Bay d'Espoir	..	65	Jewell (1939)
Salmon River	Bay d'Espoir	..	?	Jewell (1939)
North Bay River	Bay d'Espoir	..	70	Jewell (1939)
Southeast Brook	Bay d'Espoir	Milltown	65	Kemble Widmer (1950, unpub. Ph.D. thesis, Princeton Univ.)

Newfoundland after the main ice mass retreated inland.

DRIFT RIDGES: Drift ridges have been observed at the northeast end of the Burin Peninsula (D. A. Bradley, 1954, unpub. Ph.D. thesis, Univ. of Michigan, p. 105) close to the inner margin of the outer drift zone. These are as large as 1 mile long, 1000 feet wide, and 50 feet high and are oriented southeast parallel to the direction of ice movement in this region.

Bradley thought they might be unusually large eskers, but from their position amidst the general pattern of glacial deposits outlined in the present paper the writer suggests that they might be drumlins.

ESKERS, KAMES, AND KAME TERRACES: The only known eskers and kames in the outer drift zone are the ones reported by T. N. Walthier (1948, unpub. ms., Geol. Survey of Canada) at the southwest end of the Burin Peninsula. A prominent esker follows along the Fortune valley in this region (Fig. 2). Kame terraces have been reported in the Grand Bank area (Walthier, 1948) and along Salmonier Cove Pond on the Hermitage Peninsula (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 87). These features may be the products of local ice masses that remained after the main ice front had retreated inland from the coast.

OUTWASH: Numerous outwash deposits radiate from the margin of the inner drift zone outward across parts of the outer drift zone, trending coastward down river valleys and terminating in most instances as deltas at the coast. Figure 3 shows the positions of 24 outwash valley trains in eastern Newfoundland; all but five are known to end as deltas. Although these deposits at present occur in the outer drift zone, they are related chronologically to the younger deposits of the inner drift zone. Most of the outwash trains are less than 10 miles long because of the closeness of their points of origin to the coast. Those on the northern and eastern coasts are somewhat longer than their southern counterparts, probably because the lower altitudes in the north and east influenced the distance from the coast of the final ice front. Table 1 gives the locations of the outwash trains and deltas, together with references.

Outwash in the valleys of Main Brook and Shearstick Brook in the southern part of the Burin Peninsula (Van Alstine, 1948, p. 5–6) is of special interest. Here, outwash unconformably overlies stratified red clays, sands, and gravels, which Van Alstine interpreted as deltaic deposits with some interbedded till. These deposits are the only ones in eastern Newfoundland said to contain glacial material of two ages separated by a weathered till. In view of the pattern of glacial deposits shown herein (Fig. 3), further study of this area is merited.

DELTAIC SEDIMENTS: The writer observed elevated deltaic sediments at the seaward ends of outwash valley trains in Bonavista Bay and Trinity Bay and at the head of Placentia Bay. Others have been reported on both north and south coasts (Fig. 6). To date no organic material has been specifically identified in any of these deposits in eastern Newfoundland[4], although MacClintock and Twenhofel (1940, p. 1750) stated that marine fossils have been reported from a delta deposit 3 miles east of Norris Arm in the Bay of Exploits, and Coleman (1926, p. 205) reported many marine shells in a bluish till (perhaps delta bottomset clays) at Bishops Falls on the Exploits River.

The most impressive elevated deltaic sediments anywhere in eastern Newfoundland occur at Eastport and Sandy Cove in Bonavista Bay, where they are displayed as two 100-foot bluffs with extensive sandy beaches at their bases. Other deltas in Bonavista Bay are also topped at 100 feet above sea level, but deltas elsewhere around the coast of eastern Newfoundland occur at different elevations (Fig. 6).

Varved clays and silts have been observed at the base of several of these deltas, including ones at Port Blandford (Jenness, ms. in preparation), at Georges Brook near Clarenville (Hayes, 1948, p. 27), and near the mouth of the Conne River in Bay d'Espoir (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 98). The Port Blandford deposit contains about 50 varves, whereas the Conne River deposit reportedly contains more than 1600. Such bodies presumably indicate fresh-water conditions at the time of their formation and suggest that formerly large ice-blocked lakes existed in parts of Bonavista and Fortune bays and Bay d'Espoir. Fresh-water conditions would explain the apparent absence of marine shells in the deltaic sediments around the east coast of Newfoundland, in contrast to their presence in the nonvarved deltas in Notre Dame Bay and on the west coast.

Inner Drift Zone

Physiographic features.—At least two distinctive physiographic areas within the inner drift zone owe their existence to glaciation. These are the moraine-lakes area around Meta Pond and the larger area of fluted terrain a

[4] Baird (1951, p. 8), however, lists five species of marine shells found in an elevated alluvial terrace in Middle Arm of Green Bay, Notre Dame Bay, and Watson (1947, p. 3) lists one genus of pelecypod recovered from an elevated delta near the mouth of Northwest Brook in Baie Verte. Brackish-water shells are also known from the Bay St. George delta on the west coast (MacClintock and Twenhofel, 1940, p. 1753).

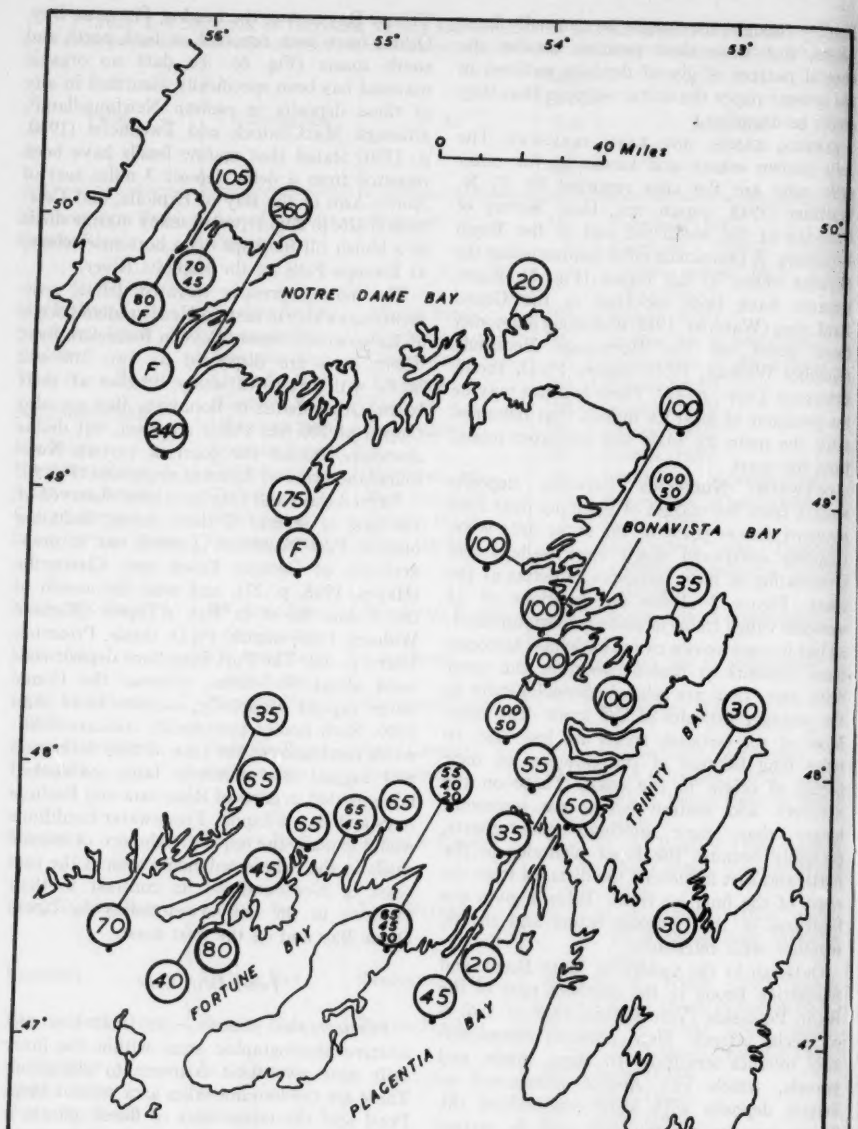

FIGURE 6.—MAP SHOWING LOCATION OF ELEVATED ALLUVIAL TERRACES IN EASTERN NEWFOUNDLAND

Elevations are in feet. Black dot beneath circle indicates delta, "F" in circle indicates shells in terrace deposits.

little farther west (Fig. 3). Other small physiographic units of this type may exist in the zone, especially west of the fluted terrain, but have to date escaped recognition. A few kettle lakes occur in the relatively little known western part of the zone.

KETTLE LAKES: The writer knows of only one occurrence of kettle lakes within the inner drift zone. This is north of Burnt Hill near the headwaters of the Northwest Gander River (J. C. Grady, 1952, unpub. map, Geol. Survey of Newfoundland). No written description of these lakes is available.

MORAINE LAKES: A multitude of small lakes, many of which trend northeast, lies within an area of approximately 400 square miles surrounding Meta Pond (latitude 48°05′N., longitude 54°50′W.). Bedrock in this area is a coarse-grained granite, with a local northeast-trending gneissic structure. The terrain has only slight relief (generally less than 50 feet) and little vegetation. Large granite boulders are scattered between the many small lakes and are found in most of the lakes themselves, indicating their shallowness. Many of the lakes are simply water-filled bedrock basins, but others appear to be enclosed by gravelly ground moraine and may be considered moraine lakes. Their northeasterly orientation evidently results from the southeast movement of the ice masses over the northeasterly foliated granitic rocks, but which of these two factors proved the greater influence is not yet known. The peculiar lake pattern in this area does not continue into the slate belt west of Meta Pond or south and southeast beyond the area underlain by granite; this suggests a close relationship between lake pattern and bedrock composition.

FLUTED TERRAIN: A few miles west of the Meta Pond moraine-lake area the land surface takes on an entirely different aspect. Here, an area of about 600 square miles between the Northwest Gander River and the ponds at the head of Bay du Nord River is characterized on areal photographs by a grooved appearance with a pronounced northwest-southeast orientation. The writer has used the term fluted terrain for this region to emphasize its physiographic distinctiveness. Striations recorded in this region (Fig. 2) indicate that the direction of ice movement here parallels that of the fluting, that is, northwest to southeast. As the area is underlain by folded argillaceous rocks, the fluting is regarded as the product of ice action upon a clayey till developed from these rocks. Presumably the final ice advance in this area scoured broad gentle troughs across the clayey till to give the strong southeast linearity it now displays.

Glacial deposits.—Most of the glacial deposits in the inner drift zone are the products of melting ice. Ground moraine is present in all parts of the zone and is separated from the ground moraine of the outer zone by an extensive, discontinuous end moraine. Many eskers lie just inside the end moraine and form a semicircular belt about 30 miles wide. Kames and kame terraces have also been recognized. Extensive outwash deposits radiating out from the margin of the inner drift zone presumably formed during the same stage of deglaciation as the deposits within the zone. Because of their geographic position, however, they are discussed under the section Outer drift zone.

The western limits of the inner drift zone are not yet known, owing to the scarcity of geological information on central Newfoundland.

GROUND MORAINE: Ground moraine is present almost everywhere within the inner drift zone, but sections indicating local thicknesses are extremely scarce, largely because most of the area is still undeveloped. The ground moraine in the inner drift zone, as in the outer drift zone, shows a strong correlation with the kind of bedrock in the immediate vicinity. This is especially true in areas underlain by granitic rocks; indeed, granite boulders and sandy debris, characteristically present in such areas, proved of considerable aid to the writer in bedrock mapping, as such accumulations seldom were found to extend for any appreciable distance beyond the margins of the granite bodies.

END MORAINE: An important, although relatively inconspicuous, glacial feature in the inner drift zone is the end moraine that lies a few miles inside the north, east, and south coasts of eastern Newfoundland. It is discontinuous, but its position can be observed in a sufficient number of places and can be inferred by related phenomena in enough other places to permit the plotting of its position with a reasonable degree of accuracy (Fig. 3). Its position is believed to mark the front of the last important stand of the ice sheet in eastern Newfoundland. The writer has observed most visible portions of this end moraine between Gander Lake and just west of Random Sound; its position from this point west along the south coast is inferred largely from the observations of D. A. Bradley (1954, unpub. Ph.D. thesis, Univ. of Michigan) and Kemble Widmer

(1950, unpub. Ph.D. thesis, Princeton Univ.). Evidence for the moraine, which lies for the most part in terrain that contains few roads and trails, includes (1) several small ridges of boulder till cut through by the Trans-Canada Highway a few miles south of Gambo, (2) boulder-strewn hillocks that occur discontinuously from south of Gambo across the east end of Maccles Pond to Terra Nova, and (3) other boulder accumulations around Salmon Pond, Andrews Pond, and Shoal Harbour Pond, south of Terra Nova.

On the south side of Gander Lake, thick vegetation, few exposures, and difficult access have made recognition of the moraine difficult. As the ice moved across argillaceous rocks in much of this area, however, it is likely that little or no moraine was ever developed in places and that the end moraine in this general area is quite discontinuous.

In the south-coast region, moraine occurs approximately where the belt of eskers and outwash deposits meet near Gisburn Lake a few miles north of Fortune Bay (D. A. Bradley, 1954, unpub. Ph.D. thesis, Univ. of Michigan, p. 104) and at the head of Bay d'Espoir (Kemble Widmer, 1950, unpub. Ph.D. thesis, Princeton Univ., p. 99–100). As the esker belt and outwash deposits are separated by the end moraine in the eastern and northern section of the inner drift zone (with which the writer is familiar), the same relationship can be expected to exist along the south coast. The position of the end moraine along the south coast has been drawn on this assumption (Fig. 3).

ESKERS: A broad belt 20–30 miles wide that contains most of the eskers in eastern Newfoundland (Fig. 3) occurs immediately inside the end moraine described. More than 50 eskers have been recognized, all trending roughly parallel to the local directions of glacial striations. The positions of most of them are shown on Figure 2. The eskers are most numerous in the granite terrain north of Fortune Bay, probably because of the ready disintegration of the coarse-grained granitic rocks in that area.

The two longest eskers in the belt can be traced for 15 miles; one trends northeast almost into Caribou Pond (south-southeast of Gander Lake), the other trends almost due south into Gisburn Lake north of Fortune Bay. Two others can be traced from 7 to 10 miles; the first follows along the private road to North Pond (west of Gambo), the second crosses Terra Nova Lake and continues along the lake to the railway and thence east.

Most of the eskers are about 10 to 40 feet high and 25 to 100 feet wide. An excellent longitudinal section of one of them is exposed on the south shore of Terra Nova Lake.

Eskers are notably absent northwest of Meta Pond, where the ice crossed a broad Paleozoic slate belt (the fluted terrain on Figure 3), and none have been observed north or east of the railway in northeastern Newfoundland.

KAMES AND KAME TERRACES: Although kames and kame terraces are commonly found in association with eskers and end moraines, very few are known to occur in the inner drift zone. Only three kames have been observed by the writer, one along the railway about a mile north of the village of Terra Nova, and two about 9 miles northeast of Meta Pond. They are low, rounded hummocks, shrub-covered, and located above or very close to granitic terrain.

Two probable kame terraces occur near the northern limits of the esker belt. One of these occurs 5½ miles south of Gander Lake on the side of a hill overlooking the junction of the east and west branches of the Southwest Gander River, more than 200 feet above the valley floor and about 400 feet above sea level. The stream that deposited it probably flowed across the ice terminus forming part of the outwash stream that followed the Gander River valley to the north. The other probable kame terrace flanks the southeast side of the ridge behind the village of Gambo. The extensive deposits occurring here, at an elevation of about 150 feet, could well have formed along the margins of a body of ice entering Freshwater Bay from the U-shaped valley of Gambo Pond. Other kame terraces probably exist within the inner drift zone but have not yet been recognized.

Glaciation on the Avalon Peninsula

The glaciation of the Avalon Peninsula is treated separately from that of the rest of eastern Newfoundland because the glacial events there differed somewhat from those farther west. Striation directions shown on Figure 2 indicate, with one exception, that ice movement throughout the Peninsula was from the higher central areas coastward. Because of the unusual configuration of the Avalon coast line, the ice masses commonly moved in opposing directions on either side of the long bays. This has certainly happened in Trinity and Conception bays and presumably also in Placentia and St. Mary's bays. It

follows therefore that the centers out of which the ice flowed on the Avalon Peninsula lay along the spines of each of the four narrow peninsulas. W. F. Summers (1949, unpub. M.S. thesis, McGill Univ.) has suggested the existence of three former centers (Fig. 7).

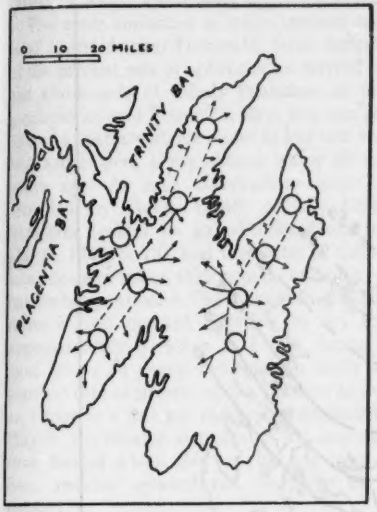

FIGURE 7.—ICE CENTERS ON THE AVALON PENINSULA
(After W. F. Summers, 1949, unpub. M.S. thesis, McGill Univ.)

The glacial deposits encountered on the Avalon Peninsula are relatively unsorted tills, invariably of local origin. These can be observed in various roadcuts, some of which have been developed for road material. Only one till has been recognized so far.

The directions of striations and the local nature of the till everywhere on the Avalon Peninsula indicate that the peninsula sustained its own ice caps and was relatively independent of the glacial activity farther west. Two recent observations provide somewhat conflicting evidence, however, and suggest that some ice from central Newfoundland may have crossed Trinity Bay. The first of these is a statement in the unpublished 1949 field notes of A. M. Christie (on file, Geol. Survey of Canada) that striations are oriented S.70°E. on a smoothly polished bedrock surface just south of Grates Point, at the northern end of the Carbonear Peninsula. If this direction has been correctly recorded, the ice moved onshore rather than offshore, in contrast to the direction of ice movement elsewhere around the Avalon Peninsula. Furthermore, the direction is almost parallel to that of the striations on the west side of Trinity Bay in Random Sound. The second observation was made just north of Heart's Delight on the east shore of Trinity Bay. Here, E. P. Henderson of the Geological Survey of Canada discovered in unsorted tills cobbles of mica paragneiss and gneissic granite which resemble rocks cropping out 30 to 60 miles to the northwest. Such rocks are unknown on the Avalon Peninsula. Presumably these cobbles were transported by ice across Trinity Bay into the Heart's Delight area. The eastern limit of the outer drift zone may therefore touch the Avalon shore at the two localities mentioned. Elsewhere it lies near the center of Trinity and Placentia bays.

A few eskers have been observed on the Avalon Peninsula; the locations and orientations are indicated on Figure 2. Sandy deposits have been reported in the valley southwest of Bay de Grave (Hutchinson, 1953, p. 27), but their mode of origin is not yet known. All other glacial deposits reported to date are relatively unstratified till.

The age of the glacial material on the Avalon Peninsula relative to the age of the deposits farther west is not known, but much valuable information could be derived from comparative studies of carbonaceous bog material in the two areas, following the manner of Livingstone and Livingstone (1958).

POSTGLACIAL COASTAL MOVEMENTS

Terraces developed on stratified deposits of gravel, sand, and clay occur above present sea level at many localities around the coast of eastern Newfoundland. Such features, indicative of fairly recent coastal emergence, have been reported by several persons (Table 1), and their locations are shown on Figure 6.

The elevations of these alluvial terraces can be summarized as follows: in Bonavista Bay they occur at 100, 50, 35, and possibly 25 and 15 feet above sea level; on the south coast (east of longitude 56°) they occur at 65–70, 55, 45, 35, 30, 25, and 10 feet above sea level; and in Placentia and Trinity bays, as well as on parts of the Avalon Peninsula, they occur between 25 and 35 feet above sea level[5].

Present knowledge does not permit ready correlation of the various levels in the different bays. The writer suggests, however, that the uppermost levels on deltas that form the sea-

[5] All figures are to the nearest 5 feet.

ward termini of outwash trains derived from the inner drift zone border are approximately contemporaneous. Although the upper surface on any one deposit does not mark the exact former sea-level position, it does mark a relative amount of uplift when compared with the upper-surface elevations on equivalent deposits elsewhere around the coast.

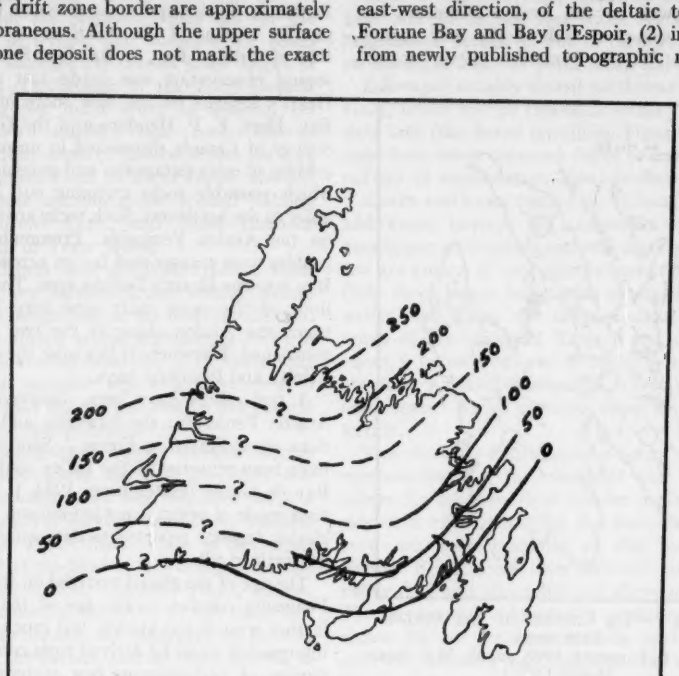

FIGURE 8.—ISOBASES IN EASTERN NEWFOUNDLAND, BASED UPON UPPER LEVELS OF OUTWASH DELTAS (Data for west coast from Flint, 1940)

Assuming that these levels represent the approximate position of a former sea level, the variation in their present altitudes would indicate differential amounts of upwarp around the east coast of the island. Figure 8 shows a set of isobase lines for Newfoundland based upon the previous assumptions, with data for the west coast from Flint (1940). These differ markedly from the isobase lines shown by Flint (1957, p. 251), particularly in the curving of the lines westward along the south coast of Newfoundland. Such a departure from the traditional northeast trend of Atlantic Coast isobase lines has been prompted by: (1) the apparent equivalence of upper levels, in an east-west direction, of the deltaic terraces in Fortune Bay and Bay d'Espoir, (2) indications from newly published topographic maps and from casual observations on shore-line features by Van Alstine (1948) and T. N. Walthier (1948, unpub. ms., Geol. Survey of Canada) that the southern part of the Burin Peninsula may be submerging at present, (3) the recent observations of G. C. Riley of the Geological Survey of Canada (1958, oral communication) that there are no indications of emerged strandline features along the south coast of Newfoundland around Burgeo, west of Bay d'Espoir, and (4) the observations of Daly (1921, p. 384) that no coastal upwarping has taken place at Port aux Basques at the southwest corner of the island. Daly partly anticipated the position of the isobases shown on Figure 8, even to the presence of the southward bulge of the zero isobase near the east coast, although he lacked sufficient data to draft a diagram. The con-

cavity of the isobases on Figure 8 toward Labrador cannot but promote renewed speculation that Newfoundland was invaded by the Labrador ice sheet, as suggested by Flint (1940, p. 1777) and others.

The coast continues to move upward northwest of the Avalon Peninsula. Some indication of the present rate of uplift can be derived from the statements of elderly fishermen at many localities around Bonavista Bay, who can point to rocks that are above water at low tide today but which were always below water 40 to 50 years ago. As such observations have been recorded by Hayes (1948, p. 13), Edward Sampson (*quoted in* an editorial footnote in Hayes, 1948, p. 13), and the writer at different localities and times, their general reliability can hardly be questioned. The rocks in each location have moved upward (relative to sea level) approximately 6 inches to 1 foot during the past 40 to 50 years, indicating a fairly substantial rate of present uplift, perhaps as much as 1 foot to 2 feet per century as suggested by Hayes. No reliable estimate of the amount of time during which this part of the coast has been moving upward can be given at the present.

Coastal movements on the Avalon Peninsula evidently differ from those to the west. A few elevated features (chiefly wave-cut terraces, but also some alluvial terraces) do occur around the coast of the peninsula, indicating a small coastal emergence at some time during the fairly recent past; however, the abundance of splendidly developed bay-head bars and spits at the heads of Trinity Bay and their absence northwest of the Isthmus of Avalon (which separates Placentia Bay from Trinity Bay) indicate that the Avalon shore line is now submerging[6]. W. F. Summers (1949, unpub. M.S. thesis, McGill Univ., p. 27) and E. P. Henderson (*quoted in* Lord, 1958, p. 21) have independently come to the same conclusion.

[6] Tide-gauge records kept at St. John's, Newfoundland, between 1936 and 1957, suggest a fairly substantial rate of rise of sea level—specifically 0.44 feet or about 2¼ feet per century. This is considerably more than the apparent rise in sea level recorded at Halifax, Nova Scotia, between 1896 and 1956 (the latest available data)—0.6 feet, or about 1 foot per century. It seems more than likely, therefore, that the coast near St. John's is sinking, perhaps at the rate of 1 foot per century, in addition to the apparent (eustatic ?) rise in sea level of the same amount. These data are perhaps the strongest evidence of present day submergence of the eastern part of Newfoundland.

The line of separation between the submerging Avalon Peninsula and the emerging coast line to the west is fairly sharp across the Isthmus of Avalon. Its approximate position is shown as the zero isobase on Figure 8.

Two factors effecting the shore line are apparently in operation around the coast of eastern Newfoundland at present; (1) an eustatic rise of sea level, believed to be about 1 foot per century, and (2) a sinking of the Avalon Peninsula together with a differential rising of the coast west of the Avalon. The number of different terrace elevations, the presence of elevated terraces around the Avalon Peninsula, and the drowned nature of much of the coast line west of the Avalon Peninsula suggest that coastal movements in the past differed considerably from those presently in force.

Summary and Conclusions

Following is a summation of late Pleistocene glacial events in eastern Newfoundland.

(1) Late Pleistocene ice from west of longitude 56° flowed across all eastern Newfoundland but stopped before reaching the Avalon Peninsula. The date of this advance has not been established. Ice masses existed on the Avalon Peninsula at about the same time.

(2) The ice front subsequently retreated inland, probably fairly rapidly, with perhaps a few minor, brief re-advances, until it reached a position well inland.

(3) The ice sheet then developed an extensive end moraine that now encircles much of eastern Newfoundland just inside the coast. This resulted either from a lengthy stand of the ice front at that position or from a final major advance to that position from farther inland. The area between this end moraine and the coast (exclusive of the Avalon Peninsula) is herein termed the outer drift zone. It is covered with unsorted ground moraine of variable thickness and relatively local derivation.

(4) Final melting of the ice sheet in Newfoundland produced the glaciofluvial deposits in back of the end moraine, in the region herein termed the inner drift zone. It also produced the outwash deposits that radiate coastward from the end moraine and terminate generally as deltas at the coast. Small valley glaciers existed in the high terrain around Fortune Bay, from which small local glacial deposits developed.

(5) Large fresh-water lakes developed at the heads of Fortune and Bonavista bays and

Bay d'Espoir, presumably as a result of ice forming blockades across the headlands of these deeply carved bays. Varved deltaic deposits accumulated where river valleys entered these lakes. The deltaic sediments around Notre Dame Bay, however, were deposited in salt water, as indicated by the presence of marine shells.

The removal of the ice cover from eastern Newfoundland has resulted in rapid upwarping in a northwest direction. This is readily recognized in coastal areas west of the Avalon Peninsula. Isobase lines drawn on the upper levels of deltaic sediments associated with outwash trains that radiate from the inner drift zone are concave toward the northwest.

Coastal movements have been more complex than outlined in this paper, but present data do not permit a fuller interpretation.

Glacial studies in Newfoundland are still in their infancy. Pollen studies on some of the many Newfoundland bogs would, however, provide much valuable information on the dates of the last glaciation and on climatic changes in the near past.

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THRUST FAULTING AND CHAOS STRUCTURE, SILURIAN HILLS, SAN BERNARDINO COUNTY, CALIFORNIA

BY DONALD H. KUPFER

ABSTRACT

In the Silurian Hills, 15 miles southeast of Death Valley, California, the dominant structure is the Riggs thrust. Beneath the thrust older Precambrian metamorphic rocks are overlain by the later Precambrian Pahrump group. Above the thrust are Paleozoic(?) carbonate rocks (Riggs formation, provisional name) and Tertiary(?) rocks, in part sedimentary and in part volcanic. Unconformably on all these bedrock units are a monolithologic carbonate megabreccia, fan gravels of two ages, and several terrace gravels. The topography is controlled by lithology and structure, drainage is subsequent, and scarps are either erosional or fault line.

The Pahrump group in the Silurian Hills is 11,000 feet thick and is subdivided into 35 mappable members, chiefly coarse clastic rocks derived from the south. Correlation with the three formations of the type Pahrump in the Kingston Range, 15 miles north, is uncertain.

Granitic rocks of two or more ages intrude the Riggs and Pahrump rocks. Distinctive members of the Pahrump group can be traced from unmetamorphosed sedimentary rocks in the west to intensely feldspathized and metamorphosed rocks in the east. The older granitic rock is displaced by the Riggs thrust fault; the younger granitic rock is both localized by the fault and displaced by it. Most of the movement on the north-trending, high-angle faults is prethrust, but contemporaneous and post-thrust movement occurred.

The Riggs thrust was apparently localized along the angular unconformity between the Riggs and Pahrump rocks. The thrust surface is anticlinal or dome-shaped, probably because of post-thrust warping. The thrust zone is a "chaos" similar to the type chaos, the Amargosa chaos described by Noble (particularly the Virgin Spring phase), except that it is composed chiefly of rocks from the footwall of the thrust instead of the plate. Detailed mapping of the chaos, possible because many of the members or beds in the Pahrump group are distinctive, shows that component fragments were carried southwestward as much as 2 miles from their original position in the autochthon and were piled up in an imbricate structure in which normal stratigraphic order is approximately maintained. The amount of movement was not determined, but the distribution of granitic rock in and under the plate suggests a minimum movement of 8 miles. A debris-flow megabreccia covers the eroded trace of the Riggs thrust fault. The Tertiary(?) volcanic rocks are folded and confined to the thrust plate, which suggests that the less intensely folded Riggs thrust is younger.

Chaos structures may have formed by the imbrication and piling up of a whole series of very small thrust plates above an underlying block, which was being shortened under compression. A few strong, competent plates moved out several miles without being broken up appreciably, but most plates broke up into giant lenses and blocks that moved much shorter distances. The incompetent material was ground up and acted as a lubricant. The stratigraphic section, thus "skeletonized" and abbreviated into "chaos", retained a crude stratigraphic order.

CONTENTS

TEXT

	Page
Introduction	182
Mapping	182
Acknowledgments	183
Geomorphology of the Silurian Hills	184
Mineral resources	187
Rock formations	187
Introduction	187
Pre-Pahrump rocks	187
Pahrump group	188
Riggs formation	194
Cenozoic deposits	196
Tertiary(?) sediments and associated volcanic rocks	196
Megabreccia	197
Fan gravels	198

Terrace gravels 199
Other late Cenozoic deposits 199
Igneous rocks 199
General 199
Diabase 199
Granitic rocks 199
Melanocratic rocks (inclusions) 200
Structural geology 200
Introduction 200
High-angle faults 201
General 201
Van fault 201
Meeker fault 201
Minor faults 202
Riggs thrust fault 202
Thrust plane 202
Riggs chaos 204
Direction and amount of displacement 206
Age 207
Folding 207
Folds in the Pahrump group 207
Folds in the Riggs formation 208
Post-thrust warping(?) 208
Intrusive pods 208
Origin of the chaos 209
Amargosa chaos 209
Riggs chaos 210
Diagrammatic presentation of origin of chaos 211
Other hypotheses 212
Conclusion 213
References cited 213

ILLUSTRATIONS

Figure Page
1. Index map of California, showing location of the Silurian Hills 183
2. Index map of portion of Death Valley-Mojave Desert region showing location of geologic mapping 184
3. Theoretical reconstruction of block-fault movement of Riggs thrust plate 185
4. Latest drainage changes in the Silurian Hills area 186
5. Graphic section of Pahrump rocks 190
6. Sketches of west wall of abandoned railroad cut in Railroad Valley, east of Middle Island 198
7. Structure contours on the Riggs thrust fault 203
8. Brunton-pace map of chaos geology in the southwestern corner of Middle Island 205
9. Complex flowage fold in carbonate rock of the Riggs formation 208
10. Pods of granitic rock 209
11. Diagrammatic sketch sections showing possible origin of chaos structure by thrust faulting 212

Plate Facing page
1. Silurian Hills, San Bernardino County, California: geologic map of western portion and structure sections 181
2. Geologic map and sections of the Silurian Hills, San Bernardino County, California 198
3. Structural features of the Silurian Hills, San Bernardino County, California 214

Introduction

Mapping

The Silurian Hills are in the northern part of San Bernardino County, along the northern border of the Mojave Desert, and 15 miles southeast of Death Valley National Monument (Figs. 1, 2) California. Large alluvial fans heading in the Shadow Mountains to the east surround the Silurian Hills and end in Silurian (Riggs) Dry Lake 2 miles to the west. The hills are topographically rugged and are characterized by steep cliffs with large talus slopes at the base. The average relief is 1500 feet. The hot dry climate supports only scanty vegetation and a negligible soil cover. Bedrock exposures are very good.

The geology was mapped on two scales. The western area (Pl. 1) was mapped on aerial photographs enlarged to a scale of 1:7200 (1 inch equals 600 feet). On completion of the field work the geology was transferred to a topographic map made by transferring contours sketched on aerial photographs to a horizontal triangulation network made from Brunton bearings. Barometric readings were used for vertical control, and some supplemental control was taken from the U. S. Geological Survey topographic map of the Avawatz Mountains quadrangle (scale: 1:250,000). No aerial photographs were available for the eastern part of the area, and the geology was mapped on the Avawatz Mountains quadrangle enlarged to 1:24,000, with contours modified in the field. The geology of the eastern area (and the western area) is shown on Plate 2. The area east of co-ordinate 10000 W. is in the U. S. Geological Survey Ivanpah quadrangle.

From 1948 to 1950, 7 months were spent in the field. Most of this time was spent in the western area.

Several unconventional symbols are used on Plate 1. The geologic contacts shown by the conventional lines are commonly fault contacts (explained below in section on Minor Faults). Both lithologic symbols (ps) and stratigraphic symbols (p_{Tr}) are used for the Pahrump rocks (explained in description of the Riggs Chaos). The zero point of the co-ordinate system is such that (20000 N.–10000 W.) is at latitude 35° 30′ and longitude 116° 00′ (Pl. 2).

PREVIOUS INVESTIGATIONS: The meager geological literature on the Silurian Hills consists mainly of brief descriptions of the mines of

the area in the annual reports of the California Division of Mines, in Mineral Resources of the United States (now Minerals Yearbook), and in a few other mining publications. The area is generalized on the reconnaissance maps of Spurr (1903, Pl. 1), Waring (1915, Pl. 2), Thompson (1929, Pl. 8), Tucker and Sampson (1931, p. 322), and Jenkins (1938). Two abstracts (Kupfer, 1951; 1953) and a brief summary report (Kupfer, 1954) have been published on the area.

FIGURE 1.—INDEX MAP OF CALIFORNIA, SHOWING LOCATION OF THE SILURIAN HILLS

Figure 2 is an index map of the hachured area. SA, San Andreas fault; G, Garlock fault; F, Furnace Creek fault; DV, Death Valley fault; B, Blackwater fault; M, Mesquite fault; P, Pinto Mountain fault; and SJ, San Jacinto fault.

Acknowledgments

A dissertation based on this field work was presented in partial fulfillment of the requirements for the degree of Doctor of Philosophy at Yale University, under the direction of Prof. C. R. Longwell. In addition to the faculty of the Geology Department at Yale University, many people examined the area and the manuscript and offered excellent advice and criticism. Among them are John Shelton, Levi Noble, and Lauren Wright who visited the area; John Sanders, James MacAllister, and Clarence Allen who read the manuscript. I wish particularly to acknowledge the assistance of D. F. Hewett of the U. S. Geological Survey, who suggested the problem, guided the field work, and through his extensive knowledge of the regional geology influenced much of my thinking.

Part of the field work was financed by a Shell Oil Company fellowship. I used the equipment and facilities of the California Institute of Technology and of the U. S. Geological Survey and was flown over the area by J. S. Shelton of

Pomona College. Mr. and Mrs. Lloyd Williams and Mrs. G. A. Meeker generously extended their hospitality.

Geomorphology of the Silurian Hills

No evidence of late Quaternary or Recent tectonic movement was found in the Silurian Hills. The present topographic form is largely the result of varying resistance to erosion among the various rock types. Carbonate rocks are most resistant and cap all the highest peaks and ridges; clastic rocks form the intermediate heights; granitic rocks form the floors of valleys.

The exposed areas of granitic rocks are mature surfaces of low relief with narrow(?) adjacent pediments. Nearby areas underlain by the more resistant Pahrump rocks, though controlled by the same base level for apparently the same period of time, are youthful, rugged surfaces with deeply intrenched canyons. Riggs carbonate rocks controlled by this same base level are still essentially undissected.

The Silurian Hills are a topographic high, elongate east-west. The northern side of the hills is a mature, fan-embayed front. Large alluvial fans extend from all the canyons, coalesce along the front, and merge with the giant alluvial fan that heads to the east. These fans have buried the mouths of the canyons and extend southward far up the canyons, locally even to the divides. The southern mountain front, in contrast, is a high, relatively straight escarpment cut by deep canyons. This escarpment formed by differential erosion, not by faulting. The mountainous area is held up by the resistant carbonate rocks of the Riggs formation, and the flat pedimentlike surface in front of it is cut on easily eroded granitic rock. The cliff face corresponds closely to the intrusive contact, which is straight because it was controlled by a straight carbonate rock-clastic rock contact. Regional drainage changes are also responsible for some of the contrast between the northern and southern sides of the Silurian Hills.

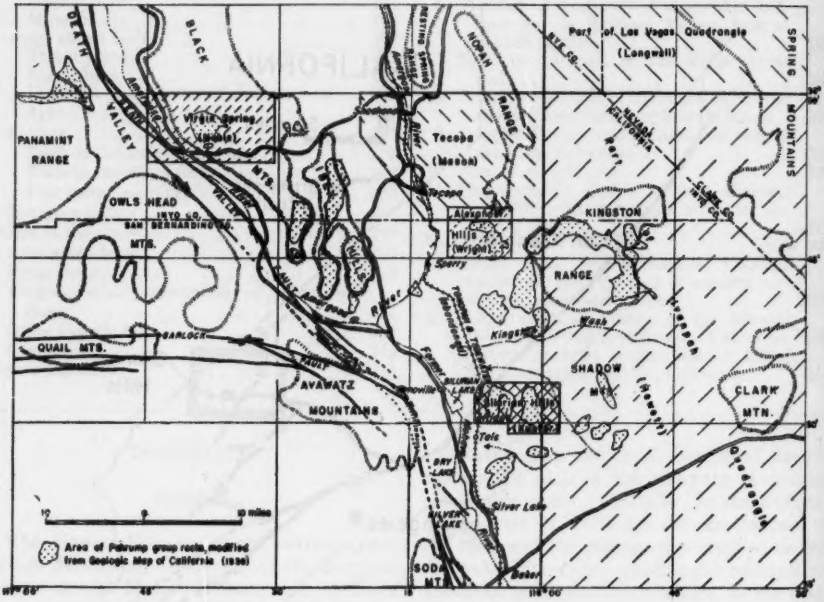


FIGURE 2.—INDEX MAP OF PORTION OF DEATH VALLEY-MOJAVE DESERT REGION SHOWING LOCATION OF GEOLOGIC MAPPING

Noble (1941), Mason (1948), Wright (1954), Longwell (1926), Hewett (1956), and Kupfer (1954; this report). All the area on this diagram north of 35° 30′ and west of 116° 00′ is covered by Noble and Wright (1954). The pattern of strike-slip faults was generalized from Noble and Wright (1954) and from Grose (1959). *See* Figure 1 for location of mapped area.

The last two important structural events in the Silurian Hills were the thrusting of a thick section of carbonate rock over the area and the slicing up of this carbonate plate by north-trending high-angle faults. Fault-guided streams cut through the cap rock into the upland formed after the last thrust faulting and "may have begun in late Pliocene time".

Blocks F and G are eastward-tilted blocks deeply eroded on the west and north but elsewhere protected by thick cap rock. Blocks H and J form a large equidimensional unit in

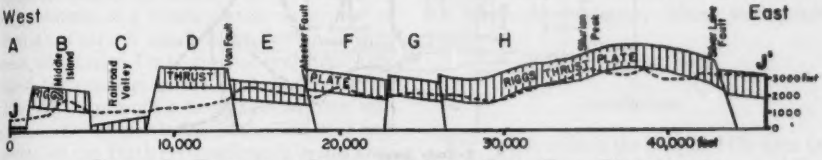


FIGURE 3.—THEORETICAL RECONSTRUCTION OF BLOCK-FAULT MOVEMENT OF RIGGS THRUST PLATE

Letters designate the individual blocks and are referred to in text. *See* Plates 2 and 3 for line of J-J'.

weaker clastic rocks below (Pahrump group) and eroded headward to form low divides. The eastward-trending hills were thus broken into a series of blocks elongate northward (Pl. 3). Thereafter each fault-bounded block was eroded independently, according to its topographic position. For purposes of description the blocks are lettered from west to east as shown in Figure 3.

Block A, west of the Silurian Hills, is down-dropped and buried under alluvium. Block B is a maturely dissected horst, the western portion of which is a pediment of unknown width. Carbonate caps held up the three peaks that are now surrounded by alluvium. Block C, a graben, is a physiographic low (Railroad Valley) composed of easily eroded granitic rock.

Block D, a horst, and block E, a tilted graben, have been inverted topographically. Block E, after downfaulting along the Van and Meeker faults, stood lower than the adjacent blocks. Its resistant cap protected it from erosion. Meanwhile the high-standing carbonate cap rock of the neighboring blocks was eroded away, allowing erosion of the less resistant clastic rocks beneath. As the neighboring blocks became topographic lows, the resistant cap rock of block E was exposed. Block E now stands as a topographic high bounded by obsequent fault scarps.

This topographic inversion may have been preceded by a stage of planation, as the upper portions of blocks B, D, and F are now surfaces of low relief. This old surface of planation(?) is at an elevation of 1500–2500 feet (highest to the east). Hewett (1956, p. 101) described a possible continuation of this surface to the east as the "Ivanpah upland". He states that which the carbonate cap rock is broadly domed. The center of the dome is breached by erosion, leaving a rim of hogbacks in quaquaversal dip away from the central core of highly metamorphosed Pahrump rock.

The geomorphic history of the Silurian Hills is also reflected in its stream patterns. The large granitic area that now extends as a pediment and buried pediment south of the Silurian Hills and south of the mapped area for an unknown distance was the first area to be reduced by erosion. Streams must have headed in what is now the southern slope of the Silurian Hills. As the granitic area was reduced to a pediment, these headwater streams increased in gradient and captured northward-flowing drainage (Fig. 4). A similar pattern of capture was occurring in the western part of the Silurian Hills as block B was reduced from a high-standing barrier ridge to the present residual peaks surrounded by alluvium. Westward-flowing streams could cross the block along new and shorter courses and capture the headwaters of northward-flowing streams.

The topographic inversion is also reflected in the stream patterns. The present upper surface of block E, between the Van and Meeker faults (Fig. 3), is physiographically mature with subdued relief and open valleys. The streams now cascade over the bounding scarps. In contrast, the streams in the easily eroded block to the west are in deep narrow canyons. Originally a stream flowed northward along the Van fault (Fig. 4A, 35000 W.). As the carbonate cap rock to the west was removed, first from the north and then progressively southward, west-flowing streams cut headward through the less resistant underlying rocks and captured the Van fault drainage (Fig. 4B). In

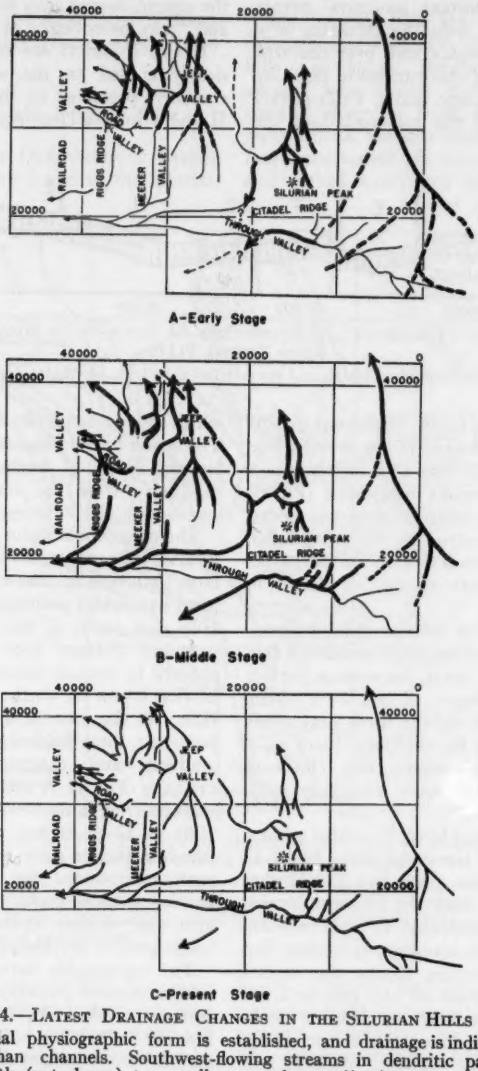


FIGURE 4.—LATEST DRAINAGE CHANGES IN THE SILURIAN HILLS AREA

Stage A. The essential physiographic form is established, and drainage is indicated by thick lines; shows trends rather than channels. Southwest-flowing streams in dendritic pattern have reduced granitic area to the south (not shown) to a pediment and are adjusting channels to a more westerly course.

Stage B represents known drainage channels that formed as west-trending streams established dominance. Channels may not be contemporaneous.

Stage C is the present drainage pattern and is also shown in A and B by thin lines for purposes of reference.

Road Valley the most southerly of the west-flowing streams made the final capture of the headwaters. The head of Road Valley is a mature, open valley, as wide as it is long, which now empties into the very narrow, high-walled, youthful canyon. Because of this increased headwater area, a large alluvial fan is now forming at the mouth of Road Valley in

Railroad Valley and is expanding over the adjacent alluvial fans. Road Valley and the several narrow canyons north of it may, however, owe part of their narrowness to a recently lowered base level as well as to stream capture.

In the Silurian Peak area, the original drainage was to the north (Fig. 4A), as shown by remnants of a broad, mature valley east of Silurian Peak. A small south-flowing stream, west of Silurian Peak, cut headward and captured the drainage of the north-flowing stream (Fig. 4B) and built a large alluvial cone west and southwest of Silurian Peak. This cone now covers all the Tertiary(?) sediments in this area. A north-flowing stream, which headed in the southeast corner of Jeep Valley, cut headward and captured this drainage (Fig. 4C), again diverting it to the north. This most recent channel is cut deep into the alluvial cone and the bedrock east of it.

Mineral Resources

Silver and talc mines have operated in the Silurian Hills, but none is being worked at present. Copper, gold, manganese, and barite also occur, but not in commercial quantities.

Primary silver occurs in galena and is localized at the intersection of small cross faults with a favorable host bed in the Riggs formation. Mineralization, spotty and irregular, is associated with dark-brown, hydrothermal (?) manganiferous dolomite and with barite. The principal production probably came from secondarily enriched oxide ores. The Riggs (Alta) silver mine may have opened as early as 1890[1] but certainly was operated from about 1913 to 1920. Tucker (1921, p. 359) reported that it produced $200,000 in silver. Several other mines and prospects may have yielded small tonnages of lead-silver ore.

Veins of barite up to 2 inches wide occur in the carbonate rocks of the Riggs formation. Barite crystals 1–2 inches in diameter have been found. Manganiferous dolomite is a common associate of barite and the lead-silver minerals.

Talc occurs in the basal Pahrump rocks, associated with diabase sills. (*see* Wright, 1950, p. 278; Wright and others, 1953, p. 200–202, for further details.) Copper mineralization was noted at several localities in the Pahrump group. Prospects in the Pahrump rocks on the northwest slope of Silurian Peak and in the general area to the north and west are said to be for gold associated with the copper.

The Silurian Hills area is not favorable for prospecting for material that requires large-scale, low-cost operation, as the rocks of the area are strongly broken up by faulting and locally metamorphosed. Moreover, the area is far from transportation, labor, water, and supplies.

Rock Formations

Introduction

The oldest rocks in the Silurian Hills area are an igneous and metamorphic complex of Precambrian age. The upper Precambrian Pahrump group, 11,000 feet thick, unconformably overlies this complex. The Riggs formation (provisional name) of late(?) Paleozoic age is thought to have been deposited unconformably on the Pahrump group, but in the Silurian Hills it now occurs only in a thrust plate that rests on the Pahrump rocks. No Mesozoic sedimentary rocks are known, but two ages of post-Paleozoic granitic rock occur. Tertiary(?) sedimentary and volcanic rocks crop out in a few areas and are overlain by partially consolidated fan gravels. Terrace gravels, taluses, wind-blown sands, slump blocks, and modern stream gravels make up the remainder of the youngest deposits.

Pre-Pahrump Rocks

The pre-Pahrump rocks crop out where block faulting and erosion have exposed the core of the Silurian Hills or as blocks in the lowest part of some of the chaotic zones associated with the thrust faulting. These rocks include metamorphosed sediments and several types of granitic rocks. The metasediments are presumed to be older than the upper Precambrian Pahrump rocks, as they appear to underlie them unconformably. Although the sedimentary contact was not observed, in several places the contact is a minor fault with no significant displacement. The age of the associated granitic rocks is less certain; in part they may be younger than the Pahrump group.

Many rocks of the complex are highly fractured and foliated and show evidence of having been subjected to two stages of metamorphism. Subschistose gneisses are the most common rock type, but true schists are present. The most common granitic rock type is a very coarse-grained granite with pink feldspars. Most of the granitic rocks are not foliated.

[1] The name Silurian Hills was probably given by a Cornish miner, after the hills of his homeland.

Pahrump Group

Hewett (1940) defined the Pahrump series and subdivided it into the Kingston Peak formation, Beck Spring dolomite, and Crystal Spring formation. The name is here changed to Pahrump group, in accordance with the rule of the Committee on Stratigraphic Nomenclature that formations are combined into groups (Ashley and others, 1933, Art. 12). Rocks of the Pahrump group are exposed throughout the Silurian Hills. They and the underlying pre-Pahrump granitic and metamorphic rocks form the autochthonous or footwall block of the Riggs thrust. Pahrump rocks also are the principal rocks involved in the chaos structure associated with the thrust. In the northwestern part of the area the Pahrump rocks are well indurated but essentially unmetamorphosed. In the eastern and southeastern part of the area they have been intruded by granitic rocks and highly metamorphosed.

In the Silurian Hills the exposed portion of the Pahrump group is more than 11,000 feet thick and can be subdivided into 35 mappable members (Fig. 5). Because of rapid facies changes in the Pahrump group, these members will probably not be recognized in other areas; hence they have not been given formal names but have been numbered from 1 at the base to 35 at the top. As will be explained, correlation of these members with Hewett's formations is uncertain. Either the 35 members are all part of the Crystal Spring formation, or they represent all three of Hewett's formations and a younger, previously unrecognized formation.

LITHOLOGY: Coarse clastic sediments dominate the Pahrump in the Silurian Hills. Beds of clay-sized particles are conspicuously absent. Sandstone, generally orthoquartzite, is the most common rock type. Beds of carbonate rock, mainly dolomite (weathering orange brown), occur throughout the section. Of the 35 members, the middle 11 are of heterogeneous lithology; the others are mainly quartzose.

The stratigraphic section of the Pahrump group (*see also* Fig. 5) was measured by pace and compass traverse in the northwest corner of the Silurian Hills and is the most nearly complete Pahrump section in the Silurian Hills area.

Members 1–12 are well-sorted quartzose or carbonate rocks; quartzite is dominant. Pebbles in the conglomerates are well rounded and consist of white vein quartz. The sand-sized material, whether alone or as a matrix in the conglomerate, is silica and forms vitreous, white or light-gray orthoquartzites. Thin dark streaks of finer-grained material in the quartzites accentuate the bedding or cross-bedding. Siltstone is rare, and shale is absent. Carbonate rock, except in member 12, occurs interbedded with quartzite as moderately thin beds of either white limestone or buff to white dolomite that weathers orange brown. Diabase sills and talc are associated with these beds.

Members 13–23 consist of clastic material of variable composition, but granitic debris predominates. Very little carbonate rock occurs in these members. Thin-bedded, rhythmically bedded sediments occur at the base and top of the interval, and coarse conglomerate or fanglomerate occurs in the middle. The conglomerate pebbles are predominantly granitic. Medium- and coarse-grained sandstones predominate in most of the section and occur in moderately thick beds (1–4 feet thick) that weather pale greenish gray, probably because of the presence of epidote. Fine-grained sandstones and coarse-grained dark-gray to black siltstones are common. The most characteristic feature of members 13–23, excepting members 18 (fanglomerate) and 20 (limestone), is their rhythmically bedded character. Several thin beds combine to form a single graded unit with coarsest material at the base and finest at the top. Some individual graded units are 15 feet thick, but many are less than 1 foot thick.

The two main conglomerates, members 17 and 18, are cobble and boulder conglomerates, and no mappable contact has been found between them. Member 17 contains well-rounded cobbles, generally granitic, and is intruded by several diabase sills. Member 18 is probably a fanglomerate, in which the cobbles are more angular, and some of the boulders are blocks 10–100 feet in diameter. Limestone, dolomite, quartzite, siltstone, and amphibolitized diabase cobbles are common. These rocks in member 18 are apparently fragments of the Pahrump group; nearby Pahrump rock must have been uplifted and eroded at this time.

Members 24–33 are well-sorted quartzose rocks of uniform lithology, with interlayered carbonate beds. The white, vitreous, massive orthoquartzite members (24, 26, 28, 33, and parts of 31) dominate the zone. They are pure and glassy, and individual grains are visible only on the weathered surface. The siltstone members (27 and 29) are dark, probably because of internal reflections in the very fine-grained constituent material rather than any significant

change in composition. The dolomite members (25 and 32) weather to a characteristic orange brown to chocolate brown.

Members 34 and 35 are apparently transitional to unexposed members higher in the section. They contain the only true shales in the sequence. The thin beds of compact, medium-gray limestone and dolomite contain algal growths similar to those that form the characteristic concentric patterns in the Beck Spring dolomite (Hewett, 1940).

THICKNESS: Hewett (1940) reported the thickness of the Pahrump at the type section as 6500 feet. Several years earlier Noble (1934, p. 2) described a generalized section, 7000 feet thick, of the "Algonkian" rocks in Death Valley. In 1941 Noble (p. 950) noted that:

"The thickness of the Pahrump series is considerably greater in the southern Death Valley region than it is in the Kingston Range, where Hewett's section was measured, and is probably nearer 10,000 than 5,000 feet."

Wright (1954) stated that in the Alexander Hills the Pahrump beds total "about 8000 feet in maximum observed thickness". The section in the Silurian Hills is 11,000 feet thick and is incomplete; the top is concealed by alluvium and megabreccia.

ORIGIN: The 35 members of the Pahrump group exposed in the Silurian Hills can best be explained as the coarse clastic facies of a series of marine transgressions and regressions. The evidence of marine origin is inconclusive. The significant features are: (1) over-all gray color, (2) much "clean" orthoquartzite, (3) clastic beds with ripple marks and cross-bedding, (4) well-bedded, well-sorted, commonly laminated sediments, (5) beds grading laterally into a thick carbonate section to the north, (6) no notable evidence of any other origin.

Member 18 looks like a nonmarine alluvial fan deposit (fanglomerate). Hazzard (1939) described a similar conglomerate in the Kingston Peak formation in the Nopah Range (20 miles north) in which the clasts are beveled and striated and probably represent a tillite, but Hewett (1940) called the same deposit in the Kingston Range a fanglomerate. Member 18 in the Silurian Hills is lithologically similar to the Kingston Peak fanglomerate and may be equivalent to it; no striated or beveled clasts were observed. Member 18 is assumed to be an alluvial-fan deposit and therefore nonmarine.

Although member 17 is gradational with member 18, the metasomatic alteration of the sills in member 17 (described below in the section on Diabase) suggests that member 17 was deposited under water. Therefore, like the rest of the Pahrump group, it is probably marine.

SOURCE OF SEDIMENTS: At several places in members 1–13 the quartzite contains large cobbles in the lee of which smaller pebbles have collected, thus preserving an indication of the direction of current flow. In view of the changing currents in marine waters, such indications of current direction may have little value, but they all suggest a northward current. Moreover, the Pahrump rocks in the Kingston Range and Ibex Hills, north of the Silurian Hills, are composed of fine-grained clastics and carbonates, whereas in the Silurian Hills they are coarse-grained clastics with negligible carbonate. This facies change combined with the evidence of direction of current flow suggests a southerly source for the sediments. In members 13–23, granitic material appears in place of the quartzose material of the lower members. This may indicate a change of source area, but no evidence of its direction was noted.

AGE: The rocks of the Silurian Hills are considered part of the Pahrump group and of late Precambrian age. Hewett (1940) described the Pahrump group in the type area as resting unconformably on the Precambrian gneissic granite and overlain unconformably by sedimentary rocks that probably represent the basal formation of the Paleozoic rocks (Lower Cambrian). Noble (1934, p. 174; 1941, p. 950) reported similar relations in Death Valley. Extension of the name Pahrump to the Silurian Hills is based mainly on lithologic correlation. The same distinctive sequence that characterizes the basal Pahrump in the Kingston Range and in the Death Valley area (Wright, 1949) occurs with minor facies changes in the Silurian Hills.

The only other stratigraphic units in the Death Valley region that might be confused with the Pahrump group are the clastic formations of the Cambrian (Hazzard, 1937). Many lithologic differences distinguish these two units. For example, the Cambrian rocks of this general region maintain about the same lithology from place to place, and shales form a conspicuous part of the sequence. The Silurian Hills Pahrump rocks, in contrast, show facies changes and, except for the top 100 feet, lack shales.

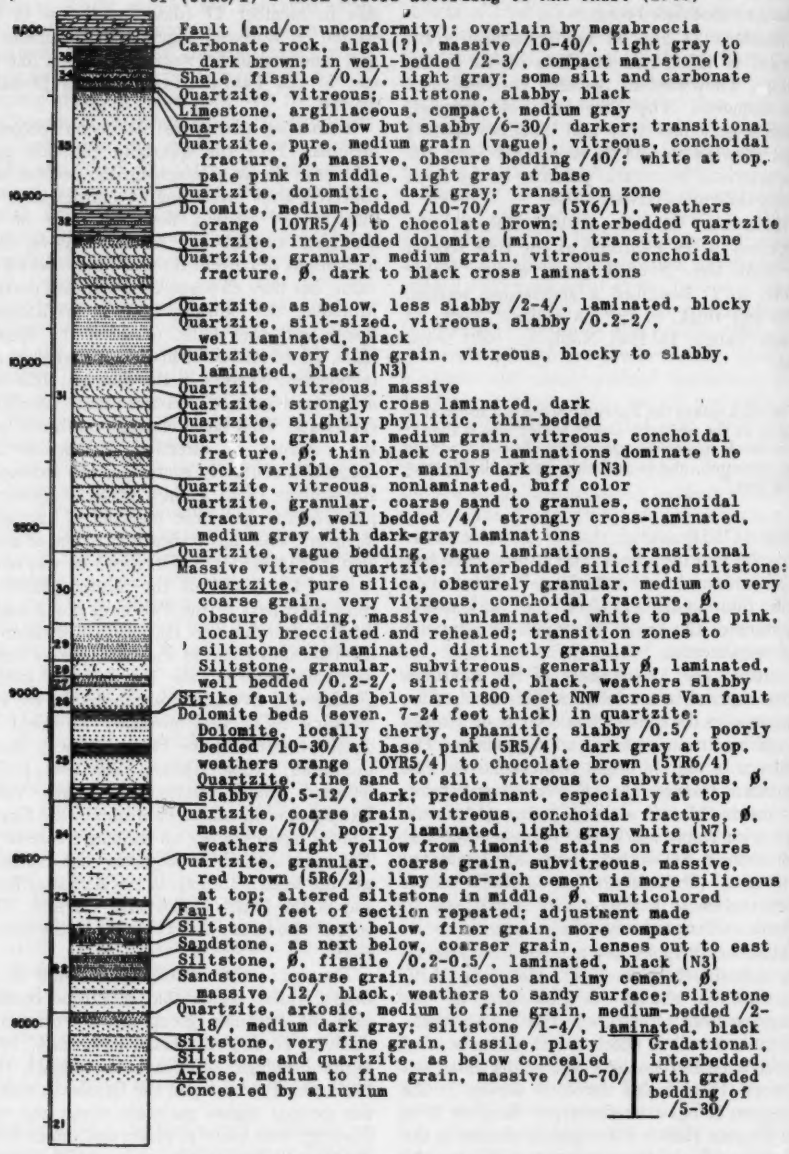

KEY: Grain size according to Wentworth; Ø = Rock breaks across the grains; /2-6/ = Thickness of bedding in inches; (N3) or (5YR6/1) = Rock colors according to NRC chart (1948)
11,000
10,500
10,000
9800
9000
8800
8000
36
34
35
32
31
30
29
28
27
26
25
24
23
22
21
Fault (and/or unconformity); overlain by megabreccia
Carbonate rock, algal(?), massive /10-40/, light gray to dark brown; in well-bedded /2-3/, compact marlstone(?)
Shale, fissile /0.1/, light gray; some silt and carbonate
Quartzite, vitreous; siltstone, slabby, black
Limestone, argillaceous, compact, medium gray
Quartzite, as below but slabby /6-30/, darker; transitional
Quartzite, pure, medium grain (vague), vitreous, conchoidal fracture, Ø, massive, obscure bedding /40/; white at top, pale pink in middle, light gray at base
Quartzite, dolomitic, dark gray; transition zone
Dolomite, medium-bedded /10-70/, gray (5Y6/1), weathers orange (10YR5/4) to chocolate brown; interbedded quartzite
Quartzite, interbedded dolomite (minor), transition zone
Quartzite, granular, medium grain, vitreous, conchoidal fracture, Ø, dark to black cross laminations
Quartzite, as below, less slabby /2-4/, laminated, blocky
Quartzite, silt-sized, vitreous, slabby /0.2-2/, well laminated, black
Quartzite, very fine grain, vitreous, blocky to slabby, laminated, black (N3)
Quartzite, vitreous, massive
Quartzite, strongly cross laminated, dark
Quartzite, slightly phyllitic, thin-bedded
Quartzite, granular, medium grain, vitreous, conchoidal fracture, Ø; thin black cross laminations dominate the rock; variable color, mainly dark gray (N3)
Quartzite, vitreous, nonlaminated, buff color
Quartzite, granular, coarse sand to granules, conchoidal fracture, Ø, well bedded /4/, strongly cross-laminated, medium gray with dark-gray laminations
Quartzite, vague bedding, vague laminations, transitional
Massive vitreous quartzite; interbedded silicified siltstone:
Quartzite, pure silica, obscurely granular, medium to very coarse grain, very vitreous, conchoidal fracture, Ø, obscure bedding, massive, unlaminated, white to pale pink, locally brecciated and rehealed; transition zones to siltstone are laminated, distinctly granular
Siltstone, granular, subvitreous, generally Ø, laminated, well bedded /0.2-2/, silicified, black, weathers slabby
Strike fault, beds below are 1800 feet NNW across Van fault
Dolomite beds (seven, 7-24 feet thick) in quartzite:
Dolomite, locally cherty, aphanitic, slabby /0.5/, poorly bedded /10-30/ at base, pink (5R5/4), dark gray at top, weathers orange (10YR5/4) to chocolate brown (5YR6/4)
Quartzite, fine sand to silt, vitreous to subvitreous, Ø, slabby /0.5-12/, dark; predominant, especially at top
Quartzite, coarse grain, vitreous, conchoidal fracture, Ø, massive /70/, poorly laminated, light gray white (N7); weathers light yellow from limonite stains on fractures
Quartzite, granular, coarse grain, subvitreous, massive, red brown (5R6/2), limy iron-rich cement is more siliceous at top; altered siltstone in middle, Ø, multicolored
Fault, 70 feet of section repeated; adjustment made
Siltstone, as next below, finer grain, more compact
Sandstone, as next below, coarser grain, lenses out to east
Siltstone, Ø, fissile /0.2-0.5/, laminated, black (N3)
Sandstone, coarse grain, siliceous and limy cement, Ø, massive /12/, black, weathers to sandy surface; siltstone
Quartzite, arkosic, medium to fine grain, medium-bedded /2-18/, medium dark gray; siltstone /1-4/, laminated, black
Siltstone, very fine grain, fissile, platy
Siltstone and quartzite, as below concealed
Arkose, medium to fine grain, massive /10-70/
Concealed by alluvium
Gradational, interbedded, with graded bedding of /5-30/

FIGURE 5.—GRAPHIC SECTION OF PAHRUMP ROCKS
Northwest corner of Silurian Hills (39400 N.–34000 W.) to (30500 N.–39800 W.)

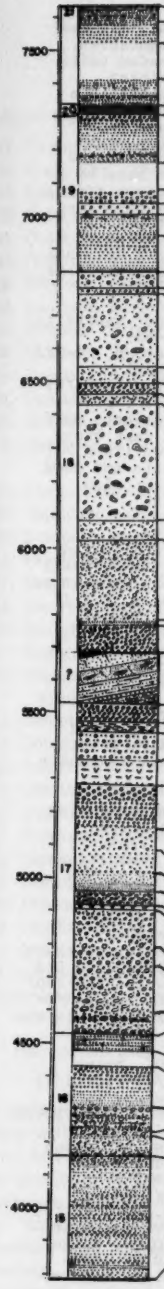
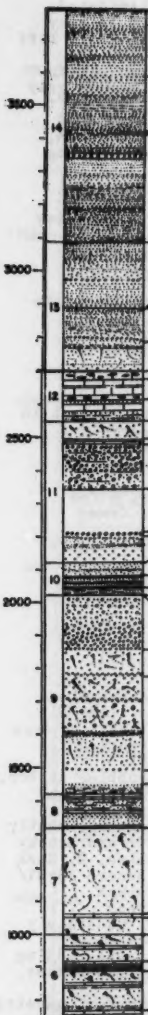

7500
7000
6500
6000
5500
5000
4500
4000
Graded beds /8/, well bedded /2-3/; siltstone, laminated, not Ø, black; quartzite, fine grain, weathers buff gray
As below concealed, but coarser grain, /8-15/, weathers buff
Concealed; probably thick-bedded /10-30/ quartzite
Graded beds /8/, well bedded /2-6/; siltstone, laminated, black; quartzite, fine to medium grain, Ø, dark blue gray
Limestone, aphanitic, slabby /0.2-3/, white, weathers buff
Quartzite, impure, fine grain, Ø, laminated, well bedded /0.1-24/ light to dark gray, weathers dark gray green; grades from fissile clay silt at base to cobbles at top
Concealed; probably mostly conglomerate
Conglomerate, large granitic cobbles, close-packed
Aplite sill, brecciated at base
Quartzite, impure, fine sand to silt, Ø, /4-8/, (N3); limy
Quartzite, arkosic, Ø, (N4) weathers (5YR5/2), graded; base, medium grain, /2-3/; top, small cobbles, /6-24/
Conglomerate, similar to next fanglomerate below
Conglomerate, numerous cobbles of amphibolite
Fanglomerate, mainly blue-black cobbles in black quartzite matrix, generally massive, some bedding near top; fragments of amphibolite, granitic rock, vein quartz, siltstone, blue-gray limestone, and buff dolomite
Fanglomerate, highly amphibolitic
Fanglomerate, typical; some boulders of buff dolomite
Fanglomerate, highly amphibolitic
Fanglomerate, mainly cobbles, but boulders up to 20 feet in diameter of varied lithology, massive, no observed bedding; fragments of black quartzite, amphibolite, siltstone, white limestone, dolomite, and others
Conglomerate, amphibolitic cobbles and boulders, a few black quartzite cobbles, amphibolitic matrix; fewer cobbles and more matrix at top
Conglomerate, very arkosic, granitic pebbles to boulders (mostly 1"-6", but up to 75" in diameter), minimum of matrix, hard, compact, very Ø, massive, unbedded, unlaminated, light-colored, weathers to buff brown
Sandstone, well bedded, blue
Conglomerate, cobble, very close-packed, as next above
Rotated blocks, well-bedded clastics and carbonates; landslumps (Precambrian or Recent ?), or chaos(?); mainly on present-day topographic highs
Conglomerate, arkosic, pebbles and cobbles, close-packed
Siltstone and claystone, hard, sinuous fracture, (5B4/1)
Conglomerate, pebble; sandstone; transitional base
Amphibolite, coarse grain, green; sill(?) or reworked amphibolite boulders in amphibolitic matrix(?)
Conglomerate, small pebbles at base to medium cobbles at top, mostly granitic, some amphibolitic(?) at top, dark-blue arkosic matrix, weathers buff or dark gray green
Arkose, coarse grain, scattered granules to boulders, mostly pebbles, Ø, good bedding /6-24/, weathers buff and blocky
Sandstone, medium to very fine grain, some pebbles, Ø, dark
Sandstone, very coarse; pebble conglomerate; obscure /8-24/
Conglomerate, small cobbles, Ø, bedded, green black
Diabase sill, partially amphibolitized, dark green
Quartzite, scattered small granitic cobbles, Ø, moderately bedded to slabby, dark bluish gray
Conglomerate, medium-sized cobbles, granitic, vague bedding
Conglomerate, large granitic cobbles, quartzite groundmass, Ø, poorly bedded, dark bluish gray
Conglomerate, granitic boulders 1-2 feet in diameter, massive
Quartzite, sand to small pebbles, Ø, unlaminated, slabby /1-4/, gray black to greenish
Concealed
Arkose, medium grain, subvitreous, Ø, well bedded; grades upward to small pebbles
Fanglomerate(?), boulders abundant, Ø, /6-30/
Conglomerate, cobbles mafic, Ø, green black
Quartzite, very coarse grain, massive /12-24/
Clastic rock, coarse grain, Ø, well bedded /2-8/, graded units of /10-40/
5-40% are very coarse grain, granules at base, epidote green (5Y6/1) to (10Y7/4)
25-50% are coarse grain, medium gray (N6)
10-70% are very fine grain, black (N3)
Member 14 (on next page) is transitional from this
Rhythmic sedimentation, graded units thicken upward in a gradational sequence; fine grain thin-bedded at base to coarse grain thick-bedded at top

FIGURE 5 (*continued*)



Sandstone, graded units common /5-24/, ø, unlaminated; 90% = very fine grain, /1-2/, (N3); 10% = medium grain, /3-12/, (10Y6/2) to (10YR4/2); many beds epidote green
Graded units as below aplite, more medium sand /3-12/; prominent gray green or epidote green beds interbedded throughout
Cobbles, granitic, 3-8 inches in diameter
Sandstone, very fine grain, granular, black (N3); gray siltstone; phyllitic sand, fine grain; quartzite, medium grain, laminated, (10Y6/2; distinct graded beds /0.1-18/, ø, interbedded epidote green beds.
Aplite sill near top of unit

Rhythmic sedimentation, graded units thicken upward in a gradational sequence; fine grain thin-bedded at base to coarse grain thick-bedded at top

Clastic sediments, rapid variations in grain size from bed to bed, fine silt to granitic boulders; mostly quartzite, fractured, ø, variably bedded, blocky, dark (N3-N5)
No granitic pebbles below this horizon
Quartzite, impure, medium grain; mafic siltstone; quartzite cobbles; all moderately bedded, ø, black (N3-N5)
Quartzite, mafic, poorly sorted (silt to pebble), ø, /4-12/
Limestone, white, locally sandy; interbedded chert
Limestone, cherty, aphanitic, well laminated, blue (5PB4/2)
Limestone, as above, nil chert, well bedded /0.3/; also limy sandstone, medium grain, /1-4/, (N5), weathers buff
Quartzite, impure, massive, vague bedding /1/, dark brown
Limestone, cherty, well bedded, platy, tan brown (10YR7/4)
Conglomerate, cobbles of white vitreous quartzite and some of dark impure quartzite, massive /6-36/, dark; also dark massive quartzite and some limy sandstone
Concealed
Conglomerate, large cobbles, a few not quartzite
Sandstone to conglomerate, well bedded /2-70/, laminated
Sandstone, limy, thin bedded /0.3-1/, slabby, transitional
Limestone, (5G6/1), weathers pale brown; as thin beds in ss.
Sandstone, limy, fine grain, well bedded, fissile /0.2-1/
No mafic or arkosic sand below this horizon
Conglomerate, pebbles to boulders of white quartzite, ø, groundmass dark vitreous laminated quartzite
Quartzite, as below but more vitreous, coarser grain, pebbles and cobbles common, well bedded /6-35/ with /18/ common, weathers blocky
Quartzite, fine to medium grain, vitreous, highly and irregularly fractured, ø, obscurely bedded; scattered pebbles and cobbles; locally laminated and dark gray; poorly exposed and yields talus slope of small fragments
Concealed
Sandstone, limy, coarse grain; with quartzite pebbles, cobbles, and pods of limestone; arenaceous marble, also with cobbles, interbedded in units 1-10 feet thick
Quartzite, medium grain, vitreous, conchoidal fracture, ø, vague bedding /8-18/, locally contains dark laminations, white to light gray brown; forms talus-covered slopes
Limestone, thin, white; sandstone, limy, with epidote
Quartzite, as next above, a few pebbles and cobbles
Quartzite, as next below, limy cement, ø, massive, weathers to granular surface; one pinkish laminated limestone bed
Limestone, cherty; quartzite; interbedded in units /10-200/
Quartzite, poorly sorted, coarse grain, ø, well bedded /3-24/, gray brown (5YR5/1); white limestone; black siltstone
Limestone, thin-bedded /0.5/, slabby, buff to white
Fault(?), possibly 30 feet lost; adjusted
Quartzite, locally phyllitic, vitreous, /10-70/, brown (5YR5/2); massive dolomite /50-100/, (10YR6/2)-(10YR5/4)
Limestone, cherty, well bedded, white; replaced by talc
Dolomite (4 feet), aphanitic, massive, weathers orange brown
Diabase, massive, fresh, sill-like
Carbonate, siliceous, conchoidal fracture, massive, green cast, weathers orange brown (5YR5/6); altered to talc
Siltstone, as below; blocky carbonate beds /24-36/
Siltstone, coarse, well bedded, slabby /0.2-6/, gray
Sandstone, pure quartz, fine grain, slabby /0.5-6/, white
Sandstone, coarse grain, fair /0.5-4/, weathers granular
Sandstone, very coarse grain, vague bedding /0.5/, white
Granule sandstone, pebbly, ø, vague bedding /0.5/, (N8)
Conglomerate, white quartz, cobble and pebble, ø, massive
Nonconformity, also slickensides
Granitic and metamorphic rock (older Precambrian)

FIGURE 5 (*continued*)

Stratigraphy of the Pahrump Group
Northwest corner of Silurian Hills

(*Cf.* Figure 5.)

	Approximate thickness (in feet)
Top not exposed	
35. Carbonate rock and shale, light gray to dark brown	60
34. Shale, silty, light gray	40
33. Quartzite, vitreous, massive, pale pink	380
32. Dolomite, weathers orange brown to chocolate brown	90
31. Quartzite, medium-grained, cross-laminated, dark gray	950
30. Quartzite, vitreous, massive, white	220
29. Siltstone, coarse-grained, subvitreous, laminated, black	110
28. Quartzite, vitreous, brecciated, slight pinkish cast	50
27. Siltstone, coarse-grained, subvitreous, laminated, black	30
26. Quartzite, vitreous, brecciated, pinkish cast	80
25. Dolomite beds, weather orange brown; in dark slabby quartzite	270
24. Quartzite, vitreous to granular, massive, gray	185
23. Quartzite, coarse-grained, red brown, lime-iron-silica cement	200
22. Quartzite and siltstone; transition from impure quartzite below to pure quartzite of members 24–33	255
21. Clastic sediments with graded bedding, mostly grayish green	700
20. Limestone, thin-bedded, slabby, white	25
19. Quartzite, impure, well bedded; conglomerate; siltstone	475
18. Fanglomerate, locally giant boulders, mixed lithology	1200
*17. Conglomerate, granitic pebbles and cobbles, arkosic matrix	900
16. Coarse-grained clastic rocks with graded bedding	370
15. Medium-grained clastic rocks with graded bedding	370
14. Fine-grained clastic rocks with graded bedding	700
13. Transition zone from quartzose (locally limy) lithology of members 1–11, to granitic lithology of members 14–19; silt to cobble sizes	390
12. Limestone, well bedded, laminated, blue	150
11. Conglomerate, quartz pebbles; dark vitreous quartzite	430
10. Sandstone and carbonate rock	100
9. Quartzite, laminated, vitreous, white; cobble conglomerate	600
8. Quartzite; limestone; limy sandstone	100
7. Quartzite, vitreous, massive, white	320
6. Limestone, white; interbedded in gray vitreous quartzite	255
*5. Dolomite, weathers orange brown; dark vitreous quartzite	160
*4. Dolomite, siliceous, limy; limestone; talc	90
*3. Phyllite; siliceous carbonate rock; fine-grained sandstone	95
2. Quartzite, fine-grained at top to coarse-grained at base	175
1. Conglomerate, white quartz pebbles and cobbles	75
Nonconformity on older metamorphic rocks	
	10,600

* Diabase sills from 10 to 200 feet thick occur in these members

CORRELATION WITH TYPE SECTION: Correlation within the Pahrump group is much less clear, but two possibilities are reasonable. The first possibility is that the Silurian Hills section represents only the Crystal Spring formation, and that the Beck Spring dolomite and Kingston Peak formation lie buried beneath the alluvium north of the Silurian Hills. This hypothesis is based principally on the facts that the Beck Spring dolomite to the north, a thick formation with a lateral extent of more than 25 miles, has no equivalent in the Silurian Hills section only 20 miles away and that the top member of the Silurian Hills section (member 35) appears to be transitional to Beck Spring type of lithology.

The second possibility is that the limestone member 12 (150 feet thick) is equivalent to the Beck Spring dolomite (1000 feet thick), and that members 19–35 represent a younger,

previously undescribed formation of the Pahrump group that is not present in the areas to the north. This hypothesis is favored by attempts to correlate transgressions and regressions in the two areas and by the presence of similar fanglomerates in the sections. This correlation would be about as follows:

Silurian Hills	Thickness (in feet)	Northern areas	Thickness (in feet)
Members 19–35	4000	Not exposed	..
Members 17–18*	2500	Upper Kingston Peak formation*	1000–2000
Members 13–16	1850	Lower Kingston Peak formation	500–1000
Member 12	150	Beck Spring dolomite	1200–1500
Members 1–11	2500	Crystal Spring formation	2000–4000

* Conglomerate and fanglomerate

Riggs Formation

NOMENCLATURE: The Riggs thrust plate is composed of carbonate rock of unknown age, for which the name Riggs formation is provisionally suggested. The stratigraphic section has not been worked out, and the structure is only partially known. The carbonate rocks are recrystallized and seemingly unfossiliferous, but evidence will be given that suggests late Paleozoic(?) age. Riggs rocks cap the major peaks and ridges and are well exposed along Riggs Ridge just north of the Riggs silver mine and in Hill 2979 just east of Riggs Ridge (Pl. 1). Gross lithologic units, primarily separating limestone from dolomite, were mapped. Many contacts were sketched in from a distance; no detailed mapping was attempted. Much of the limestone has been dolomitized, so that this lithologic mapping may have little stratigraphic significance.

If future work should establish an unmetamorphosed correlative of the Riggs rocks with a more nearly complete and structurally less complicated section, the name Riggs should not be considered as having priority over any name given to the new section (Ashley and others, 1933, Arts. 21–22). The stratigraphy of the Riggs rocks has not been worked out, a type section has not been established, the limits of the unit are unknown, and the age has not been determined.

LITHOLOGY: The Riggs formation consists of interbedded recrystallized limestone and dolomite with a few thin beds of orthoquartzite and other clastic rocks. From a distance the massive limestone units (Pl. 1, rl) appear to be blue, but they are medium gray, N5 (National Research Council, 1948), and weather to a little darker gray. The less massive, bedded limestones (Pl. 1, rl) are light gray to white and commonly weather dead white.

Fresh dolomite is white, pink, or brown but characteristically weathers to buff (about 10YR7/4). It occurs in thick, massive beds of uniform color (Pl. 1, rd) or interbedded with the white limestone described above (Pl. 1, rm). The massive units and many of the thinner units are brecciated into fragments half an inch to 3 inches in diameter and are recemented with their own debris. Dolomite is very brittle and probably was brecciated during orogenic deformation, particularly folding.

Stratigraphic sections of the Riggs formation were measured at several places; thickness ranges from 1000 to 2200 feet. These figures are of little help in estimating the true thickness of the formation, however, because the structure is not understood in detail and because the formation is locally dolomitized. The sections may not give a minimum thickness, as the area is highly faulted, and repetitions could occur. Several of the sections could fit together to give a composite section many thousands of feet thick. My best estimate is that the Riggs formation is about 2500 feet thick and that the following section, the most detailed observed, is typical.

AGE AND CORRELATION: No fossils have been found in the Riggs formation, and very little is known about its stratigraphic position. As explained below in the section on the Riggs Thrust Fault, it may have been unconformable on the Pahrump group before thrusting. It is unconformably overlain by Tertiary(?) rocks.

The Riggs formation may represent a portion of the geologic column hitherto unidentified in this region; but until some evidence confirms this, it seems best to try correlation with some

Detailed Section of the Riggs Formation from Peak 2979 to Meeker Valley

Detailed section down the ridge on the east side of Peak 2979 to the falls in Meeker Vallley. The top 50 feet of the section occurs west of Peak 2979. Unless otherwise noted, the colors are of the weathered rock, the characteristic color of the outcrop.

Fault	*Thickness in feet*	
A. Buff and blue limestone, interbedded	50	70
Buff and white limestone, interbedded, well bedded	20	
B. Massive blue limestone, laminated		145
C. Buff dolomite, weathering light buff (10YR7/4), white where fresh, thin-bedded, well bedded, fine-grained, compact	55	65
Buff cherty dolomite	10	
D. Clastic sediments, thin-bedded, intruded by granitic rock		33
E. Buff cherty dolomite		35
F. Buff and white limestone, interbedded		12
G. Massive blue limestone, well bedded		85
H. Cherty limestone, well laminated, well bedded, some dolomite; blue at top, buff in middle, white at base		90
I. Buff dolomite, cherty, well laminated		100
Fault. Probably minor amount of section lost		
J. Buff dolomite		440
Very cherty buff dolomite, massive (10YR6/2)	110	
Buff (10YR8/2) dolomite; white where fresh, cherty, limy, bedded	140	
Buff (10YR6/4) dolomite; cherty, limy, fair bedding	50	
Buff (10YR6/2) dolomite; gray where fresh, limy, massive, brecciated	140	
K. Blue limestone, lead-silver mineralization occurs in this unit		125
Blue and white limestone, interbedded, well laminated	60	
White limestone	5	
Massive blue (N7) limestone, blue (N6) where fresh, well laminated	60	
L. Buff limestone and blue limestone		85
Buff, well bedded, well laminated, lenses of blue	50	
Massive blue	15	
Blue and white, interbedded, well bedded	20	
M. Buff dolomite, buff to white where fresh, some brecciation		90
N. Blue-gray limestone		175
Blue and white limestone	15	
Blue limestone and buff-weathering pink dolomite, some chert	20	
Massive blue limestone, gray where fresh	30	
Buff dolomite and white limestone, interbedded, well bedded	20	
Gray (N5) limestone, well laminated	50	
Buff-weathering pink dolomite and blue limestone, interbedded	40	
O. Gray limestone		170
Massive limestone, dark buff gray, brecciated	21	
Massive gray limestone, some laminations	20	
Buff (5YR6/2) dolomite, tan brown where fresh, fair bedding, chert	11	
Massive blue limestone, gray where fresh, well laminated	18	
Thin-bedded limestone, buff, gray, and white	50	
Massive gray limestone (5PB7/2) where fresh, some laminations	50	
Note: End of section. Beds below estimated from adjacent areas		
P. Buff dolomite, brecciated		230
Q. Blue limestone		50
R. Buff dolomite		200
Total		2200

part of the known column. Correlations considered are: Beck Spring dolomite, Noonday dolomite, upper Paleozoic carbonate rocks, post-Paleozoic rocks.

As pointed out in the section above on Pahrump Correlations, the Beck Spring formation may not be represented in the Silurian Hills but may lie to the north under alluvial cover. In the section on Structural Geology it is shown that the Riggs thrust plate arose from about this same area to the north. The Riggs formation, however, is composed chiefly of well-bedded blue limestones and light-buff dolomites, whereas the Beck Spring dolomite is massive, dark-gray dolomite. It is unlikely that the two are correlative.

All the Cambrian rocks of the general region have been described by Hazzard (1937). The only significant carbonate unit is the Noonday dolomite, 1500 feet thick; it occurs at the base of 10,000 feet of Lower Cambrian clastic sedimentary rocks. The Riggs formation does not resemble the algae-rich Noonday dolomite, which over a wide area has a uniform lithology and is characteristically light buff on weathered surfaces. Paleozoic rocks of eastern California and southern Nevada have been described by Hewett (1931), Noble (1934), Hazzard (1937), Nolan (1943), Mason (1948), and others. Except for the basal 10,000–11,000 feet of dominantly clastic rocks mentioned above, most of the Paleozoic section consists of carbonate rocks. The Riggs formation is probably the correlative of part of this upper Paleozoic section.

Post-Paleozoic sedimentary rocks in this general region, in so far as they have been recognized, are generally clastic or volcanic. I know of no post-Paleozoic unit similar to the Riggs formation.

Fossil plant fragments were found in a wedge of carbonate rock in a fault zone (21500 N.–20400 W.) that separates rocks of the Riggs formation from Tertiary(?) sediments. No other rocks crop out in the immediate vicinity, and the displacement on the fault is probably not large. The carbonate rock does not look like the Riggs rocks in the immediate vicinity, but it does appear similar to some manganiferous dolomites in the Riggs about 1500 feet away. Thus, these fossil plants do not necessarily come from the Riggs formation, but that is their most logical source. The fossil material was examined by Roland Brown, U. S. National Museum, who identified the material as probably *Equisetum*(?). This is not a diagnostic fossil, but the *Equisetum* were most abundant in the Carboniferous.

In summary, the Riggs formation is younger than the Crystal Spring formation (late Precambrian) and older than the Tertiary(?) sediments. The best evidence suggests that it is late Paleozoic.

Cenozoic Deposits

Tertiary(?) sediments and associated volcanic rocks.—The Cenozoic deposits of the Silurian Hills and their interrelationships are as follows:

A. Recent alluvium, taluses, and landslide blocks
B. Four or more levels of terrace gravels
(Change in external drainage conditions)
C. White fan gravels
(Minor erosional unconformity)
D. Pink fan gravels
(Angular unconformity—formation of megabreccia)
E. Tilted white gravels
(Angular unconformity—probable formation of chaos)
F. Folded brown to white gravels and conglomerates
G. Folded purple and red sandstones
H. Volcanic rocks
(Angular unconformity)

Units E, F, G, and H constitute the Tertiary(?) rocks and were mapped as a single unit. Units A, B, C, and D are not differentiated on Plate 1, and units C and D are not differentiated on Plate 2.

Folded sandstones, semiconsolidated gravels, and associated volcanic rocks occur in the northeast corner of the Silurian Hills and at a few other places in the mapped area. No fossils were found in the sediments, and their Tertiary(?) age is merely suggested. Probably these sediments were deposited at several different times, and at least one unconformity has been found within them. No stratigraphic section was determined. The volcanic rocks are principally lapilli tuffs of about latite composition (megascopic examination) and are light purplish, reddish, or brown. Most of the folded purple and red sandstones are poorly consolidated and crumbly, but some are well cemented with calcite. The purplish-gray (10YR6/2 to 5GY6/1) sandstones contain abundant volcanic material, especially tuff. The red and orange (10R5/2 to 10R6/6) sandstones are generally more quartzose. Higher in the section the grain size increases, and granules and pebbles become common. The folded brown to white gravels and conglomerates are loosely consolidated and poorly bedded. The fragments consist of

Pahrump rock, granitic rock, gneiss, and minor amounts of volcanic rock. In a few places boulders 10 feet or more in diameter are a minor though conspicuous part of the unit. The tilted white gravels are very loosely consolidated, but locally a few beds are well cemented. Pebbles and small cobbles predominate, and most are granitic and gneissic.

The age of the Tertiary(?) sediments is critical in the determination of the age of the Riggs thrust fault and the associated granitic rocks. Tertiary is the conventional age for sediments associated with unconsolidated volcanic materials in the Mojave Desert area (Hewett, 1954, p. 15), but these rocks may even be Cretaceous.

Megabreccia.—An essentially monolithologic megabreccia of Goodsprings(?)[2] dolomite crops out along the extreme northern border of the Silurian Hills. The main mass of the megabreccia is composed of a medium-gray dolomite intensely brecciated and recemented with its own carbonate. Bedding is almost obscured. The fragments or blocks are generally a few inches to several feet in diameter and angular, but a few are 50–100 feet long. Fragments of tetracorals, brachiopods, crinoid brachials, and *Syringopora* were found in the blocks and identified by C. O. Dunbar as anywhere from Ordovician to Pennsylvanian.

In the northeast corner of the area, the more southerly outcrops of the megabreccia (along co-ordinate 32000 N.) are more typically sedimentary (less tectonic) in appearance than the main mass and may have been, in part, water-laid. Fragments larger than 6 feet are rare. Many of the smaller fragments are rounded. The fragments consist of both blue and black carbonate rock and some quartzite; indistinct bedding is observed in the matrix.

Megabreccia is similar to the chaos material described below and locally gradational to it. Typical megabreccia, however, is distinct in many ways. The fragments are smaller, more angular, and more brecciated. Fine-grained matrix material is common and in some places is clastic rather than gougelike. Megabreccia blocks show evidence of rotation, but very few show the distinct surfaces of movement so common on chaos blocks. Finally, megabreccia is commonly interbedded with or closely associated with gravels and other alluvial debris, whereas typical chaos is not.

The megabreccia in the Silurian Hills thickens eastward from a thin deposit interbedded in gravel (Fig. 6) to a widespread, massive deposit split into two parts. The lower (southern) breccia contains smaller fragments than the typical megabreccia, and some fragments are rounded. In a few places the matrix appears to be Tertiary(?) fine-grained clastic debris. The upper breccia is separated from the lower by a wedge of Tertiary(?) gravels and has the more characteristic tectonic aspect. About 1–2 miles east, across an alluvium-filled valley, the megabreccia again occurs (Hewett, 1956, p. 88–90). Here the "dolomite breccia-conglomerate" or "giant breccia" as Hewett calls it is much thicker, and some individual fragments are very large. He describes (p. 90) one block that is 1000 feet long and 300 feet thick. Looking toward this area (south half of Hewett's Figure 34) from the Silurian Hills, the "giant breccia" looks like a thick-bedded unit of the Paleozoic bedrock.

Similar megabreccias have been described by Henshaw (1939, p. 6), Lamey (1948, p. 46–47), and Jahns and Engel (1949) about 15 miles southwest of the Silurian Hills, at the south end of the Avawatz Mountains. Grose (1959), who has mapped the Soda Mountains still farther south, also finds megabreccia.

The Silurian Hills megabreccia appears to be transitional between a tectonic and a sedimentary breccia. Gravity controlled its movement, but faulting must have produced the topographic features that set up the conditions favorable to gravity sliding. In addition, as noted in the section on the Origin of Chaos, strike-slip faulting and/or thrust faulting may have caused the original brecciation. Longwell (1951, p. 354), in describing similar megabreccia near Hoover Dam, said:

> ". . . differences between a breccia of this type and the chaos may in some circumstances be gradational. A thrust block above a fault with moderate or low inclination presumably would supply fragments of exceptional size, would aid gravity in propelling the debris downslope, and conceivably might progressively override the debris, to form a chaos sheet of large dimensions."

In describing the Ivanpah megabreccia, Hewett (1956) states:

> "Obviously extraordinary processes have contributed to deposit lenses of such material in the midst of ordinary clastic sedimentary rock, largely clay, sandstone, and conglomerate. (p. 90)
>
> "It is also a plausible assumption that there is some relation between the coarse breccia-conglomerate . . . and this thrust plate. . . . Such materials could be deposited in a sedimentary basin that lay east of the plate and the plate could later move over its own waste in that basin." (p. 99)

However, "The fatal weakness of this assumption is that there is widespread evidence that the middle

[2] Used in the sense of Hewett (1931, p. 13)—Upper Cambrian to Devonian

Tertiary sedimentary rocks were tilted and eroded before the plate of pre-Cambrian gneiss moved over them." (p. 99)

On the geologic maps and sections the megabreccia is shown as underlain by a fault, because wherever observed the lower contact was a surface of movement. Some geologists may object to this usage of the word fault.

In the Silurian Hills, the megabreccia is contemporaneous with the Tertiary(?) sediments with which it is interlayered (Fig. 6). It is younger than the Riggs thrust fault, whose eroded trace it crosses. The megabreccia is overlain by, and therefore older than, the pink fan gravels. If the megabreccia is associated with thrusting, as the above authors suggest, this is a younger episode of thrusting than the Riggs thrust.

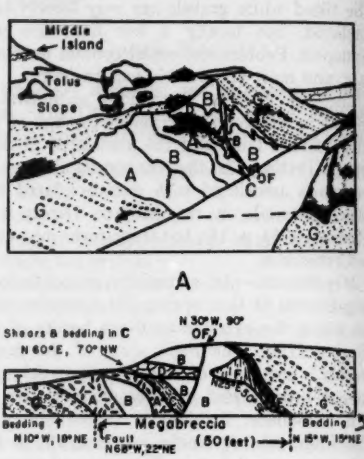
Noble (Noble and Wright, 1954, p. 152) described the Jubilee phase of the Amargosa chaos as similar to megabreccia and noted that it is younger than his more typical chaos, the Virgin Spring, from which it appears "to be separated by an unconformity of erosion":

"The Jubilee phase, which overlies both the Virgin Spring and Calico phases, is in part a mosaic of huge irregular fault blocks of brecciated granite (,) and of volcanic and sedimentary rocks of Tertiary age, and in part a series of sedimentary rocks (chiefly fanglomerate) interlayered with megabreccia. . . . [Noble] now believes that they may in part represent debris-flow accumulations . . . [as] described by Longwell (1951). . . . the Jubilee phase of chaos locally constitutes a basal part of the Pliocene-Pleistocene(?) Funeral fanglomerate and passes upward into the Funeral through a series of interlayered megabreccia and fanglomerate beds."

All the best described megabreccias occur in areas of low-angle thrust faulting but are interbedded with sediments younger than the well-exposed thrusts. With the possible exception of the thrust described by Hewett (1956, p. 95–98), no contemporaneous or younger thrust has been observed.

Fan gravels.—Pink fan gravels occur along the northeast border of the Silurian Hills and locally are cemented to a fanglomerate. They have a high content of volcanic material, which on weathering colors the rock pink to red brown. The gravels are poorly exposed because they weather easily to rounded, reddish, rock-strewn surfaces that yield no outcrops. These surfaces cover most of the extreme northeast corner of the area (Pl. 2) and conceal the relations between the underlying megabreccia and the Tertiary(?) sediments. These pink fan gravels are younger than the megabreccia and the Tertiary(?) sediments, older than the white fan gravels.

White fan gravels occur mainly in Jeep Valley, where they are by far the most abundant



A

B

FIGURE 6.—SKETCHES OF WEST WALL OF ABANDONED RAILROAD CUT IN RAILROAD VALLEY, EAST OF MIDDLE ISLAND

A is a perspective view of the cut, and B is a sketch to scale. The megabreccia (B, also A, C, D), about 10–15 feet thick, is interlayered with tilted gravel (G). Both are capped by alluvial and talus gravel (T) and dip away from the observer at about 15°. The megabreccia, generally recemented into massive rock, is here sheared and deformed. In some places it is highly sheared, altered, and stained yellow green (A); in others it is highly faulted and sheared but unaltered (C); in others it is only broken into angular blocks (D) (34400 N.–43200 W.). A—altered breccia; B—breccia; C—sheared breccia; D—blocky breccia; E—gouge; F—fault face, slickensided; G—gravel, tilted; OF—open fracture; T—talus gravel.

rock type. The gravels are poorly indurated by carbonate cement and include fragments of all bedrock types exposed in the area. They lie unconformably on the pink fan gravels in the southeast corner of Jeep Valley (28000 N.–22000 W.) and are cut by the highest terrace levels (30000 N.–28000 W.). The gravels have an initial dip of 1°–10° and locally have slumped to produce dips as high as 25°. There is no evidence of tilting.

A deposit of gravel on the south slope of the Silurian Hills (Pl. 2), on a ridge 170 feet above the main Riggs Wash (21400 N.–22000 W.), is thought to be a correlative of the white fan gravels. In its present position 170 feet above

the stream channels and terrace deposits, it indicates that during deposition of the white fan gravels a large alluvial fan spread out southward, as well as northward, from the Silurian Hills. Except for this one outcrop, later erosion has removed the southern fan, and present drainage is adjusted to a much lower base level.

Terrace gravels.—Terrace gravels occur in all the major canyons. Several terrace levels (maximum of four) occur in each canyon. Terrace levels could not be correlated from canyon to canyon as if they were controlled by an external change of base level. They probably reflect local controls such as drainage changes from stream capture and minor course changes due to a gradually lowering local base level.

The terrace deposits are moderately well stratified. Any large vertical exposure shows a great variation in grain size, but the individual layers are much more nearly uniform; each bed formed from fragments of a single size range. For example, one bed 6 inches thick may contain pebbles 1–3 inches in diameter and a minimum of finer-grained materials, and the next bed 3 inches thick may be all coarse-grained sand. A few beds contain boulders and cobbles, but the smaller sizes greatly predominate.

The terrace gravels were studied at 11 localities to determine composition, particle size, and roundness. In composition the terrace gravels reflect the rock types exposed farther up the canyon. Rock fragments tend to remain angular during transportation down canyon but decrease rapidly in size. This is particularly noticeable for the limestone fragments, and as they are soluble in water it suggests rapid movement down canyon during cloudbursts and flash floods. Between storms the channels are dry and the fragments unchanged.

Other late Cenozoic deposits.—The youngest deposits in the area are alluvial gravels, wind-blown sands, taluses, and slump blocks. The gravels occur in the stream courses and on the alluvial fans that surround the area. Wind-blown sands are most abundant in Railroad Valley, where they are thin but extensive. Taluses occur below all the steeper cliffs and scattered over most of the lower slopes. Land-slump blocks are associated with many of the high cliffs. Most cliffs consist of a cap rock of massive resistant carbonate rock (Riggs formation of the thrust plate) resting on brecciated and friable clastic rock (Pahrump group in the Riggs chaos), a relationship particularly favorable to land slumping.

Igneous Rocks

General.—Igneous rocks of several ages appear in the Silurian Hills. The oldest are associated with the pre-Pahrump metamorphic rocks and were not studied separately. Diabase sills of late Precambrian age occur in the Pahrump group. Mesozoic(?) granitic rock, mainly quartz monzonite porphyry, is the most extensive. Aplite dikes and minor pegmatite segregations cut the quartz monzonite porphyry and may be of the same intrusive episode. Podlike masses of medium-grained granodiorite are the youngest.

Diabase.—Diabase sills occur in several members of the Pahrump group. The association of the lower sills with the basal members of the group (basal Crystal Spring formation) and with talc has already been reported (Wright, 1949; 1950, p. 278). In the Silurian Hills the lower sills are slightly transgressive; they occur in member 4 at the Annex and Berryhill talc mines on the west, at the apparent top of member 4 at the Van talc mine (relations obscured by faulting), and at the base of member 6 at the Ceramic mine on the east. These lower sills are partially altered to amphibolite.

Mafic rock, originally of diabasic composition but now almost completely amphibolitized, occurs as sills in member 17 and as sedimentary fragments in member 18. Intrusion must have occurred at the time of deposition of the upper beds of member 17 or the lower beds of member 18. Amphibolitization was probably caused by connate waters in the sediments. The intensity of alteration increases upward in the section and probably is related to depth of burial at the time of intrusion. Members 4–6, which were deeply buried and presumably well compacted, contained little water, and the diabase sills were only slightly altered. Sills injected near the surface, into the very wet sediments of member 17, were highly altered. Some of the mafic rock may have been extruded onto the floor of the sea.

Granitic rocks.—The most abundant granitic rock in the area is pinkish, coarse-grained biotite-quartz monzonite porphyry. Megascopically it is characterized by pink orthoclase phenocrysts, half an inch in diameter, imbedded in a white or gray groundmass of feldspar and quartz. Biotite, the only prominent mafic mineral, is not abundant. The rock is transitional between granite and quartz monzonite, as the potassium feldspar is about twice as abundant as the plagioclase. This is the

Teutonia quartz monzonite of Hewett (1956, p. 61–67). Miller (1946, p. 500, 522–523) reported a chemical analysis, norm, and mode for a similar rock near Baker, 20 miles south of the Silurian Hills.

This quartz monzonite porphyry, as well as some less common equigranular rocks, intrudes the Riggs and Pahrump formations and occurs as pebbles in the Tertiary(?) sediments. Both porphyry and sediments are cut by the Riggs thrust. Nothing else is known of the age of the porphyry in the Silurian Hills. In the desert region quartz monzonite and associated granitic rocks are commonly assigned to the late Cretaceous or early Tertiary (Hewett, 1931, p. 38; Noble, 1941, p. 954; Miller, 1946, p. 508), but any age from Middle Jurassic on is possible (Longwell, 1950, p. 428; Hewett, 1954, p. 12).

The other important granitic rock of the area is a medium-grained, equigranular, gray quartz monzonite or granodiorite. Megascopically it is nonporphyritic and finer-grained than the porphyry; plagioclase is generally dominant. In a few outcrops orthoclase is more abundant than plagioclase. The rock is leucocratic but may contain up to 1 per cent biotite. Except for one large mass just south of the mapped area, the equigranular rock occurs in small pods. These pods are described in the section on structure and are shown to be younger than the quartz monzonite porphyry and contemporaneous with the movement on the Riggs thrust.

Small masses of other granitic rock types, such as granite and quartz diorite, were not mapped separately, and their age relationships were not determined. I mapped several aplite dikes (Pl. 1) that cut the quartz monzonite porphyry and the Pahrump group. Minor lenses (segregations?) of pegmatite a few inches wide and a few feet long occur in the quartz monzonite porphyry but were not mapped.

Melanocratic rocks (inclusions).—Melanocratic rocks of unknown age are abundant in the quartz monzonite porphyry and have been interpreted as inclusions. They range from fragments a fraction of an inch in diameter to masses measurable in hundreds of feet. The larger inclusions are shown separately on Plate 1. Almost all are feldspathic (not carbonate).

The main mass of quartz monzonite porphyry is south of the Silurian Hills, and its northern contact is a generally straight line along the southern border of the hills from just south of Riggs Ridge on the west to Through Canyon on the east (Pl. 3). The inclusions in this mass are very abundant close to the contact and appear to have been derived from clastic sediments. The present contact, between quartz monzonite and Riggs carbonate rock, must correspond very closely to a former contact between carbonate rock and clastic rock. The quartz monzonite must have selectively intruded or formed from the clastic rock but not from the carbonate rock.

Structural Geology

Introduction

The Silurian Hills are a small group of rugged peaks and ridges surrounded by a vast piedmont alluvial fan. The peaks and ridges are capped by carbonate rocks of the Riggs formation that have been thrust south-southwestward over all the older rocks. The folded and faulted Riggs thrust surface is exceptionally well exposed and has been mapped in detail (Fig. 7; Pl. 3). Associated with the thrust surface and equally well exposed is a complex of broken, faulted, and brecciated rocks of the chaos type. This Riggs chaos is composed mainly of footwall rocks—the well-indurated, partially metamorphosed, clastic rocks of the Pahrump group—but includes some other types. The chaos grades downward into relatively undeformed rocks of the Pahrump group.

High-angle faulting, folding, and igneous intrusion occurred before, during, and after the thrust faulting, but maximum activity occurred before the thrusting.[3] Post-thrust movement on the high-angle faults broke the thrust plate into a series of tilted fault blocks that dominate the present topography.

Igneous intrusion, *lit-par-lit* injection, and feldspathization of the older rocks are common. Below the Riggs thrust distinctive members of the Pahrump group can be traced, with minor interruptions, from relatively unmetamorphosed sedimentary rocks in the west to intensely feldspathized and intruded rocks, called "Archean" by earlier workers, in the east. In parts of the eastern area the rock has been so feldspathized that more than 50 per cent of the rock is granitic—none of it in units of mappable size. Carbonate units are generally less feldspathized and can be mapped with greater precision across the area.

Above the thrust all the carbonate rocks have been recrystallized; the most intensely recrystallized and intruded rocks occur in the west-

[3] Wright and Troxel (1954, p. 43) seem to imply, incorrectly, that all the high-angle faulting postdates thrusting.

central part of the area, where several large masses of granitic rock were mapped (Pl. 1).

High-Angle Faults

General.—Faulting seems to have been continuous in the Silurian Hills during much of post-Paleozoic time; most of the faults are high angle and trend N. 10° E. The movement was essentially vertical, continuous over a long period, and measurable in thousands of feet. Some high-angle faulting continued after the Riggs thrusting and after the formation of the megabreccia, but no evidence of late Quaternary or Recent faulting has been found.

All the faults tend to be less conspicuous (die out?) toward the south, but this is mainly because the amount of post-thrust movement is small; to the south the thrust plate covers the area. Probably the true effect is that the faults die out upward in the thrust plate; faulting to the south at depth under the thrust should be as great as in the exposed areas to the north.

Of the eight major high-angle faults in the area, only the Van and Meeker faults are well exposed. Four others are partially exposed, and two are hypothetical. These faults have broken the rocks of the Silurian Hills into a series of blocks (Fig. 3) that are reflected in the stratigraphy, structure, and topography. Measureable displacement on the faults is 1000–3000 feet of normal separation, but at least one, the Meeker fault, is in part reverse. Dips of three of the faults are unknown and of two others are not proved. The area was folded and tilted after faulting, and the classification of some of the faults as normal and some as reverse is based upon their present chance attitude, which conceals their essential unity. Therefore, they are referred to merely as high-angle faults.

The faults are described from west to east. Reference should be made to the diagrams showing the lettering of the fault blocks (Figs. 3, 7) and to the structure map (Pl. 3). The westermost fault is hypothetical; presumably it limits the western extent of the pediment under the alluvium. The next fault (between blocks B and C) is also buried under alluvium; it separates the Riggs-Pahrump rocks on the west from granitic rocks on the east. A small segment of the fault, between Pahrump rocks and megabreccia, is exposed along the northeast border of Middle Island. The third fault, between blocks C and D, is exposed only at its extreme south end, but its position is well established between outcrops; it bends sharply from N. 30° W. to N. 20° E. Both the second and third faults are unusual in that they show large post-megabreccia movements. The Van and Meeker faults, fourth and fifth in this sequence, are well exposed; they bound the downdropped block E containing peak 2979 and are considered in more detail below. The sixth and seventh faults (F-G and G-H) are well exposed on the south side of the Silurian Hills but are concealed under the fan gravels and alluvium of Jeep Valley on the north.

The evidence for the eighth or eastern fault, the Silver Lake fault, is mainly from south of the area of Plate 2. Continuity of the fault across the map area is hypothetical; possibly two or more discontinuous faults were incorrectly joined to make the one large fault shown. I believe this fault extends 4 miles south of the map area to the Silver Lake talc mine.

Van fault.—The Van fault, with an average trend of N. 5° E. and an eastward dip, is the best-exposed high-angle fault in the area and also has the clearest topographic expression. Erosion removed most of the upthrown block, and the fault is now obsequent. Where Pahrump rock occurs on both sides of the fault the net slip, assuming vertical displacement, is 2700 feet (this includes movement of all ages). Where the Riggs thrust occurs on both sides of the fault, the post-thrust displacement is apparently only 200–500 feet. To the north the megabreccia is undisplaced by the Van fault. To the south the fault dies out, but 200 feet of post-thrust displacement was measured on a lesser fault to the west. This fault may have taken up the movement.

The Van fault appears to flatten upward as its dips range from 70° in the lowest exposures to 30° in the highest ones. Actually, these flat dips occur where the steeply dipping Van fault and the flatly dipping Riggs thrust curve in together and meet tangentially (34000 N.–35000 W.). The resulting fault zone shows the characteristics of both faults (Pl. 1, sections B-B′, K-K′). The Van fault formed first and displaced the Pahrump beds 2000 feet; then the Riggs thrust formed and, in part, accommodated itself to the surface of the Van fault. During the thrusting, movements were occurring on both the Riggs and Van faults, first a few feet on one, then a few feet on the other. Thus one fault did not displace the other, but both accommodated themselves to the same surface. After thrusting, the Van fault was displaced another 200 feet. No faulting occurred after formation of the megabreccia.

Meeker fault.—The Meeker fault east of block E is moderately well exposed and oc-

cupies a subsequent (fault-guided) valley. At its north end it splits into two faults of about equal displacement. The Pahrump rocks are displaced 2500 feet vertically (west side down); this is the total net displacement. The Riggs thrust surface is displaced 500 feet or more (west side down), and the megabreccia about 400 feet (west side up).

The simplest explanation of this 400-foot post-thrust, postmegabreccia reversal of movement is a hinge movement. Assume that the 500-foot displacement of the Riggs thrust occurred after the megabreccia was formed; the west side of the block moved down at the south end, where this displacement was measured. At the same time the west side was moving up at the north end where the megabreccia displacement was measured; this upward movement is a reversal of the earlier displacements. If this assumption is correct, the main post-thrust movement on the Meeker fault postdated the megabreccia, whereas on the Van fault it predated the megabreccia.

Minor faults.—The western part of the Silurian Hills is cut by a maze of faults. On Plate 1 the major faults are shown by the conventional fault symbol. Minor faults of small displacements, fault contacts within the chaos, sedimentary contacts, and igneous contacts are all shown by the conventional symbol for geologic contacts (thin line).

The minor faults fall into several categories. Those bounding chaos blocks are the most numerous. Minor high-angle faults associated with and subparallel to the major high-angle faults are common. A third group, flat faults associated with thrusting, has displacements that range from just perceptible to almost as large as those of the major faults.

The fourth group are strike faults. Recognition of these faults is important in determining the proper stratigraphic succession. One of these strike faults, very prominent and continuous, occurs west of the Van fault and in the heart of the type Pahrump section of the Silurian Hills. It has cut out members 27-30 and part of members 26 and 31. Another strike fault, in the northwest corner of the area (North Island), has cut out members 27 and 28. In the northeast corner of the area (36000 N–17000 W.) a fault has cut out members 26 and 27. These were the only prominent strike faults observed, and all three occur at the same position in the section. Possibly all three were the same fault, a very old fault that predated the north-trending high-angle faults. A more reasonable explanation is that strike faults like these are common, but only those in the upper part of the Pahrump section, away from the chaos structure, were recognized.

A compilation of the strike and dip of the remaining minor faults shows them to be grouped around three attitudes: (1) strike N. 70°–80° E., dip 50°–60°NW.; (2) strike N. 35°–55° E., dip 60°–70° NW.; and (3) strike N. 38°–40° W., dip 55°–75° NE.

The steep escarpment on the south side of the Silurian Hills that resembles a fault scarp is the result of differential erosion along the contact between the resistant Riggs carbonate rock and the nonresistant granitic rock. No persistent fault could be traced along the southern front of the Silurian Hills, but several short faults with westward trend were located. If several of these faults acted in echelon pattern, some minor displacement of the topography may have taken place, but not enough to account for the height of the escarpment.

Riggs Thrust Fault

Thrust plane.—The Riggs thrust fault is well exposed throughout the Silurian Hills area (Pl. 3). Because the folds and faults of the Riggs thrust surface are partially reflected in the latest formed folds and faults of both plate and autochthon, it is assumed that the thrust surface was essentially planar at the time of thrusting, and that its present irregular shape (Fig. 7) resulted mainly from post-thrust deformation.

Noble (1941, p. 971) noted that the Amargosa thrust fault and the associated thrust faults in the vicinity of Death Valley generally occur on a "major unconformity between physically unlike rocks"; such a contact is a favorable location for a major break. This also seems to be true for the Riggs thrust, as the original Riggs-Pahrump contact may have been an unconformity. The evidence is as follows:

(1) The Pahrump beds dip almost at right angles to the thrust surface, whereas the Riggs beds are commonly subparallel to the thrust surface. A reviewer of this paper pointed out that much of the bedding in the Riggs formation is shown at right angles to the thrust surface in sections A-A′, E-E′, and others. In most of these cases the thrust surface is at considerable depth below the surface, and probably the surface attitudes are extrapolated too deep. The

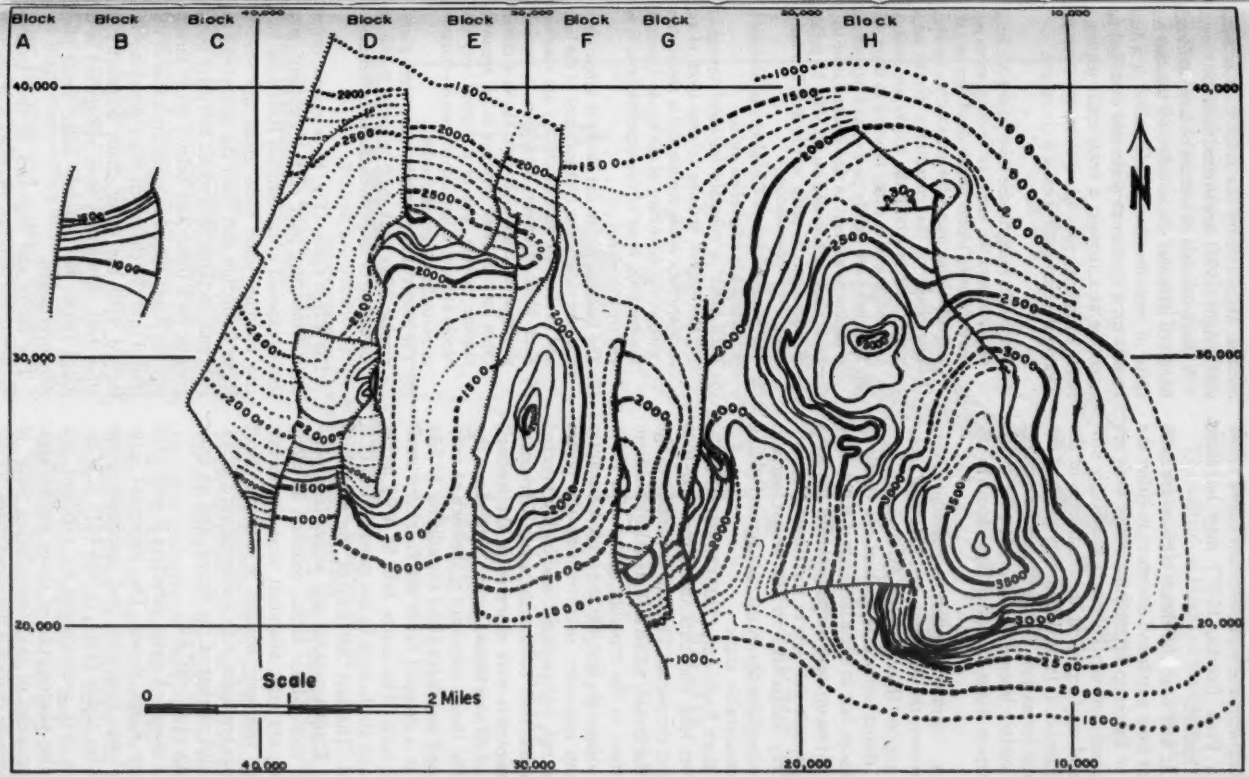


FIGURE 7.—STRUCTURE CONTOURS ON THE RIGGS THRUST FAULT

Contour interval 100 feet; contours are dashed where approximate and dotted where inferred. High-angle faults, indicated by hachures on down side, are shown only where displacement of the thrust is large. Block letters at top refer to Figure 3.

more diagrammatic representation shown under Silurian Peak (section H-H′) may be more nearly correct.

(2) At and near the thrust surface are beds different from any found elsewhere in either the Riggs or Pahrump formations. These beds are transitional between Pahrump and Riggs lithology and may represent the basal beds of the Riggs formation, formed from the debris of the Pahrump formation. The beds occur along the thrust surface irrespective of what part of the Pahrump or Riggs formation is adjacent to the fault.

(3) A "marked angular unconformity" occurs between the Pahrump and Paleozoic rocks in this region (Hazzard, 1937, p. 301).

The thrust surface is a clean-cut break in most places, but in a few localities it is not clear which of two or more major breaks is the main one. Although the beds in the thrust plate are commonly subparallel to the thrust surface, at some localities the thrust cuts across bedding at a considerable angle.

The Rock Eagle adit (24000 N.–35000 W.) is 1300 feet long and cuts the thrust 900–1000 feet from the portal. The adit is parallel to the dip of the thrust, which is 9° as calculated from the surface to the exposure in the adit. From portal to face, the first 900 feet is predominantly Pahrump quartzite; massive near the portal but sheared, feldspathized, and highly injected by granitic rock toward the fault. Carbonate rock fragments were observed in the footwall of a crushed zone 90 feet wide, 135 feet from the fault. The thrust comes out of the back at a dip of 21° and undulates along the adit for 95 feet before it is dropped below the floor on a small normal fault. The gouge zone is 6–24 inches thick, and all the rock in the general vicinity is crushed. The last 300 feet of the adit is in bedded Riggs limestone and secondarily dolomitized rock. Minor faults, common all along the adit, are most abundant near the thrust; they are relatively random in attitude. Most numerous are faults with northwestward strike, subparallel to that of the thrust, and an antithetic dip of 45°–60° SW.

Riggs chaos.—A chaos structure is associated with the Riggs thrust. The structural term "chaos" was originated and comprehensively defined by Noble (1941, p. 963–977) in a descriptive text accompanied by excellent and instructive photographs of chaos assemblages. His description of the Virgin Spring phase of the Amargosa chaos fits the Riggs chaos in most details. The principal features of a chaos, modified somewhat by the evidence found in the Silurian Hills and by the later work of Noble and Wright (1954), are summarized:

A chaos consists of a mass of large and small blocks of irregular shape collected together in a state of semidisorder. Noble called it a fault breccia on a cyclopean scale and described the structure as a mosaic, a crazy-quilt pattern, a rude shingling, or an imbricate structure. Among the characteristics of a chaos are:

(1) The blocks range in size from pods a few feet in diameter and too small to be mapped on ordinary scales to blocks hundreds of feet on a side and containing mappable units within the block.
(2) The shape of the constituent blocks ranges widely, but generally the blocks are lenticular and elongate, with either tapering or blunt ends.
(3) The blocks are tightly packed and separated by a minimum amount of fine-grained material.
(4) Each block is bounded on all sides by surfaces of movement; in other words each is a faulted block.
(5) The arrangement of the blocks is confused in detail but has an over-all order.
(6) Each block is minutely fractured throughout, yet the original bedding in each block of sedimentary rock is discernible and is sharply truncated at the boundary of the block. Commonly the bedding, even of incompetent beds, is not greatly disordered.
(7) The bedding within the blocks is subparallel to the long dimensions (flat surfaces) of the blocks and to the fault with which they are associated.
(8) Each block represents rock from a particular stratigraphic interval. Several blocks of different types may be piled together in an apparently random fashion, but in a general way these blocks are in stratigraphic sequence—commonly in normal stratigraphic order.
(9) The blocks of any one chaos represent a stratigraphic range that may be short (within one formation) or long (Precambrian to Tertiary).

The Riggs chaos (Fig. 8) is almost entirely rocks from that part of the Pahrump group that is exposed in the Silurian Hills area, but it also includes some rocks from the overlying Riggs formation and from the underlying pre-Pahrump metamorphic rocks. Intrusive pods of granitic rock (described below) occur throughout the Riggs chaos, in some instances as if intruded into the chaos but more commonly as if faulted with the other chaos fragments. No other rock formations were recognized in the Riggs chaos, but possibly some small blocks were overlooked.

The faults and fractures of the Amargosa chaos "do not dislocate the surface of the underlying gneiss" of the footwall (Noble, 1941, p. 978). The Amargosa chaos rests on the thrust

surface and consists entirely of the allochthonous material from above the thrust surface. The Riggs chaos lies under the thrust surface and consists dominantly of material from the footwall. This difference is important, as it helps to refute the argument of those (Sears, 1953) who have suggested a landslide or gravity-controlled mechanism for chaos assemblages based on their seeming similarity to megabreccia. Landslide or landslump masses can be highly fractured and disturbed, but very few, if any, strongly deform the underlying bedrock. The Riggs chaos occurs under the plate or presumed landslide and consists dominantly of footwall materials, and brecciation also extends downward with diminishing intensity 500–1000 feet into the footwall. It seems unlikely that this could be the result of gravity sliding.

The individual blocks of the chaos have been mapped and, where possible, correlated with the relatively undisturbed section of Pahrump rocks in the northwest corner of the Silurian Hills. Certain blocks were identified as to rock type, but their stratigraphic position was not determined. In Plate 1 these blocks have a special letter symbol and color to designate lithology. To find the formational members with which a block might be correlative compare its color with the colors assigned to each member. In Plate 2 several members are grouped under a single map symbol; in some places this generalization allows stratigraphic identification where none is shown in Plate 1.

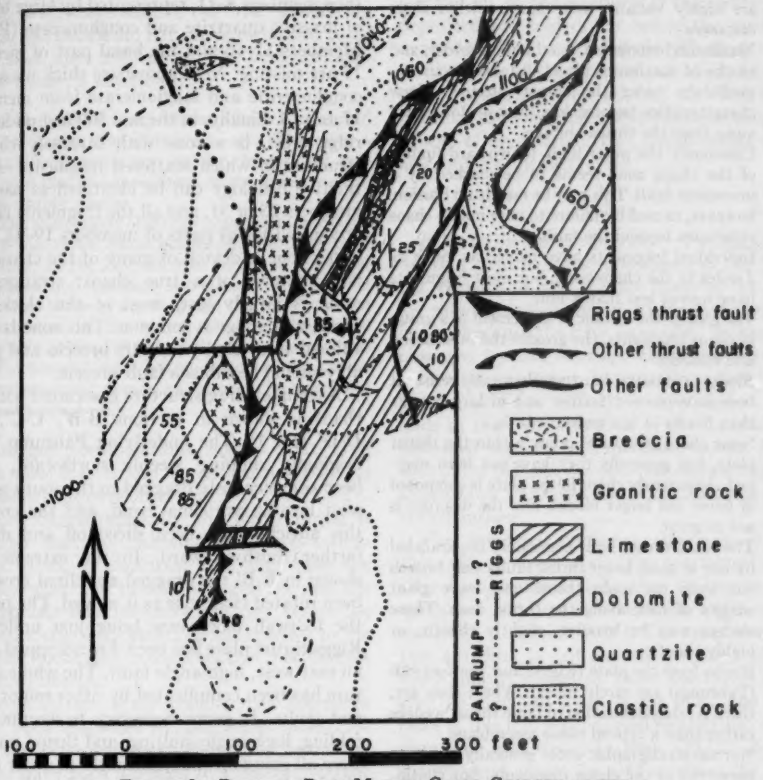


FIGURE 8.—BRUNTON–PACE MAP OF CHAOS GEOLOGY IN THE SOUTHWESTERN CORNER OF MIDDLE ISLAND

The Riggs thrust fault, a second large thrust, and several smaller thrusts have faulted the Pahrump rocks and Riggs rocks together into a chaos.

Detailed mapping of the Riggs chaos reveals the following features:

(1) The thickness of the chaos zone, the degree of slicing and disorder within the zone, and the apparent travel distance of chaos fragments are highly variable and change within short distances.
(2) Maximum brecciation, maximum disorder, and blocks of maximum travel distance occur immediately under the thrust surface. These characteristics become less intense downward away from the thrust surface.
(3) Commonly the outer limit (downward limit?) of the chaos zone seems to be marked by a prominent fault. This may be merely an illusion, however, caused by failure to note minor chaos structures beyond the fault.
(4) Individual fragments have moved as much as 2 miles in the chaos zone, but most fragments have moved less than a mile.
(5) The greater the distanc of movement of a group of chaos fragments, the greater the brecciation and disorder.
(6) Blocks of massive, hard, mechanically resistant beds have moved farther and in larger units than blocks of less resistant beds.
(7) Some chaos structures occur within the thrust plate, but generally they have not been mapped. Apparently chaos in the plate is composed of fewer but larger blocks, and the disorder is not so great.
(8) The main thrust fault commonly is paralleled by one or more lesser thrust faults that branch out from the main thrust and leave giant wedges of rock along the thrust zone. These wedges may be massive, slightly chaotic, or highly chaotic.
(9) Blocks from the plate (Riggs) and the footwall (Pahrump) are rarely mixed. Where they are, there is interpenetration of large thrust wedges rather than a typical chaos assemblage.
(10) Normal stratigraphic order generally has been preserved in the chaos fragments, but continued faulting, particularly the wedge type of faulting, has caused local repetition and inversion of the stratigraphic sequence.
(11) A normal stratigraphic section several thousand feet thick may be represented in the chaos zone by blocks that total a few hundred feet thick.
(12) Pahrump blocks in the chaos have been moved generally southward or southwestward. Each chaos fragment has a source in the autochthon that lies to the northeast. No fragment occurs in the chaos at a point north or east of its only possible source in the autochthon.

The best exposure of typical chaos structure is along the west side of Riggs Ridge where the section of Pahrump rocks, originally 11,000 feet thick, has been sheared and thinned down into 1500 feet of outcrop. In spite of the deformation most of the 35 members can be recognized. The older metamorphic rocks occur at the base of the ridge and are in part chaotic. Next east and above them are the basal members of the Pahrump group (members 1–6) and then members 8–11, represented by large blocks of massive quartzite and conglomerate (Pl. 2). Members 12–16 and the basal part of member 17 are missing. Next above are thick masses of conglomerate and fanglomerate from members 17 and 18. Finally, in the last 500 feet under the ridge crest, is a zone with intensely chaotic structure in which scattered fragments of distinctive lithology can be identified as parts of members 23 or 31, and all the fragments can be accounted for as parts of members 19–33. This upper zone is devoid of many of the characteristic features of a true chaos: stratigraphic order is largely lost, most of the blocks are small, and gouge is common. This zone is transitional from chaos to fault breccia and probably should be called a fault breccia.

A complex drag structure associated with the chaos is shown in sections B-B′, C-C′, and D-D′ (Pl. 1). The underlying Pahrump beds, originally dipping steeply northward, have been conspicuously dragged to the south-southwest into an anticlinal bend, and the crest of this anticline has been sliced off and moved farther southwestward. In the extreme case shown in B-B′ the dragged anticlinal crest has been rotated clockwise as it moved. The part of the footwall block now lying just under the Riggs thrust plate has been downdropped along an east-west, high-angle fault. The whole structure has been complicated by other minor folds and faults. It seems necessary to assume that folding, high-angle faulting, and thrust faulting were all going on simultaneously in order to explain this structural feature. Folds like these are not typical "chaos" but are transitional between the normal drag effects associated with faulting and typical chaos.

Direction and amount of displacement.—The following lines of evidence indicate that the plate of the Riggs thrust has moved south to southwest relative to the underlying autochthonous block:

(1) All fragments in the Riggs chaos have been derived from the north or northeast and carried to the southwest.
(2) Drag structure under the thrust plate is bent to the southwest.
(3) Members 12–16 are missing from the chaos under Riggs Ridge. The thrust plate in moving south-southwest to Riggs Ridge passed over the

Van fault at the point where it cuts out members 12–16.

(4) Flat thrusts of minor displacement generally show movement to the southwest.

Assuming that the Riggs thrust plate moved south-southwest relative to the underlying block, the maximum net slip is estimated to be not more than 5–10 miles. The evidence, suggestive only, is as follows:

(1) Individual blocks in the chaos have been moved 1 or 2 miles from their source area. This gives a minimum movement on the thrusts and suggests that the movement of the plate was appreciably greater.

(2) The main center of intrusion and metamorphism in the thrust plate occurs just north of peak 2979. In the footwall the degree of metamorphism and intrusion increases to the east. It would be necessary to return the main center of intrusion of the thrust plate northeastward about 5–10 miles to bring it into an area where the footwall is, by extrapolation of the known metamorphic trends, as intensely intruded and metamorphosed.

Strength considerations suggest that even 5–10 miles may be too far for a thrust plate to be pushed out over other rock.

Age.—Stratigraphically the Riggs thrust is younger than the Pahrump and Riggs formations, which it displaces, and older than the megabreccia, the tilted white gravels, and the fan gravels that lie unconformably across the eroded trace of the fault. Its age relative to the granitic rocks and to the Tertiary(?) sediments is not so clear.

Large masses of the granitic rock appear to have been displaced by the Riggs thrust. At any one place the rocks above and below the thrust have been intruded to a different extent and metamorphosed to a different degree. Thus, the fault is younger than much of the granitic rock. But granitic "pods" occur along the thrust faults in both intrusive and fault relationships to the surrounding rock. "Pods" are best explained as the result of intrusion that occurred at about the same time as thrusting. The Riggs thrust is younger than the main granitic intrusion, but it must be contemporaneous with either the final phases of intrusion or with some later intrusion—probably the latter.

Tertiary(?) sediments older than the tilted white gravel were observed only on the thrust plate. They are not found across the eroded edge of the thrust or on Pahrump rocks. They are more tightly folded than the underlying Riggs thrust plate. The simplest explanation of these structures is that the Tertiary(?) sediments were deposited on the Riggs formation and folded before Riggs thrusting.

In summary, the best evidence indicates that the Riggs thrusting occurred (1) after most, but not all, of the Tertiary(?) sediments were deposited and folded, but before the youngest were deposited and warped; (2) after most, but not all, of the igneous intrusion; and (3) after most, but not all, of the high-angle faulting. In the complex post-Paleozoic orogenic history of the Silurian Hills, the thrusting was a late, but not the last, event.

Folding

Folds in the Pahrump group.—Folding in the Silurian Hills is not so prominent as the faulting, perhaps in part because many of the folds are minor features related to the faulting, and in part because the folds have been broken and obscured by the faulting. Folds were present before the thrusting, but prethrust and interthrust folds are difficult to distinguish. The folds in the Pahrump rock of the footwall are described separately from those in the Riggs rock of the plate. A possible post-thrust warping folded both Pahrump and Riggs rocks and the thrust surface.

The principal prethrust structure in the Pahrump group is a slightly overturned anticline with an axial plane that strikes west to northwest and dips at a moderate angle to the south (Pl. 1, section B-B′). The exact shape of this anticline is not known, as most of the south limb is concealed under the thrust plate and the alluvium south of the hills. The axial trace of this fold was not well located and is not shown on the structure map (Pl. 3), but it should trend across the area in about the position of the 27000 N. co-ordinate. The beds north of the trace generally strike N. 55° W. and dip 60°–90° NE. with local overturning. South of the trace the beds would be essentially horizontal except that they have been modified by several minor north-trending synclines and anticlines. Two of these are indicated on Riggs Ridge (Pl. 3) and shown somewhat diagrammatically in section K-K′ (Pl. 1) by the bedding in the Pahrump group and the foliation in the metamorphic rocks. The minor folds are strongly disturbed by the high-angle faulting (to which they may be related) and are largely inferred from incomplete exposures in the valleys along the south border of the hills. The talc zone crops out in five places along the southern front from extreme west to extreme

east and was the best marker for this interpretation.

Folds in the Riggs formation.—The Riggs carbonate beds are rather complexly folded; many folds are of the incompetent or flowage type typical of carbonate rocks (Fig. 9). Because

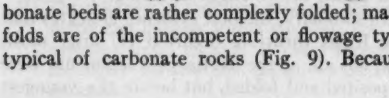


FIGURE 9.—COMPLEX FLOWAGE FOLD IN CARBONATE ROCK OF THE RIGGS FORMATION

Middle Island (34300 N.–45200 W.). Rock to the left is granitic.

the stratigraphic section was not determined and because most of the Riggs formation was not mapped in detail, the structure of the formation has not been clearly ascertained. Two observed folds that may be representative are described below. These and other folds are indicated in Plates 1 and 3.

The largest fold is the overturned syncline under peak 2979 (Pl. 1, sections E-E′ and K-K′). The sequence of beds on the east slope of peak 2979 has been interpreted as the lower limb of a syncline, and the faulted and incomplete section on the west slope as the overturned upper limb. In the structure sections the beds have been numbered to show this interpretation.

A small isoclinally folded recumbent syncline is well exposed in the cap rock of Middle Island. Its complexity may be typical of many of the other less well-exposed folds. The axial plane dips eastward at a low angle and is broadly synclinal (southeast plunge) as a result of the warping of the whole recumbent fold into a southeast-trending trough. The fold is further complicated by flowage folds (Fig. 9) and by drag and faulting related to thrusting. The most probable interpretation of this fold is shown in sections A-A′ and J-J″ (Pl. 1).

In a general way, the beds of the Riggs formation have been tightly folded into overturned synclines and anticlines and much deformed by flowage. The larger folds are complicated by minor folds on the limbs. Flowage folds generally form at some depth (under load), and the thrust probably formed near the surface, so that these folds are probably older than the thrusting and unrelated to it.

Post-thrust warping(?).—The inconclusive evidence for post-thrust warping consists of the warped shape of the thrust surface, of compatible folds in the overlying and underlying rocks, and of the orientation of the folds with respect to the direction of thrust movement.

In the western part of the Silurian Hills the thrust surface is warped into a broad anticline that trends N. 50°–55° W. across the area and is broken somewhat by faulting (Pl. 2). The crest of the anticline is generally flat, and the axis dips gently northwest at the northwest end and steeply southeast at the southeast end. Comparable structures are found in the Riggs formation. It is not easy to prove that this warp affected the footwall Pahrump rocks, but this is suggested by the attitude of the bedding north of Riggs Ridge (in the area between coordinates 35000 W. and 40000 W.); elsewhere the Pahrump beds are too poorly exposed or too broken into chaos to be interpreted. According to the pre-thrust Pahrump structure, the bedding north of Riggs Ridge was essentially vertical and parallel to the axial plane of the proposed warp. If warping occurred after thrusting, the vertical beds on the south limb would have been tilted to the north and given a "normal" dip. The beds on the north limb would have been tilted to the south and overturned. This is the present attitude of the Pahrump beds in that area.

In the eastern part of the area the thrust surface and plate are warped into a broad dome with the center just southeast of Silurian Peak. This dome is shown by the structure contours on the Riggs thrust (Fig. 7) and the quaquaversal dip of the hogbacks of the Riggs formation. The effect of the dome on the footwall could not be determined, because the footwall rocks in this area are highly metamorphosed.

Another line of reasoning suggests post-thrust warping. If thrusting was along an already undulatory surface, the ribs and troughs should be more like grooves on a fault and roughly parallel to the direction of movement. If thrusting and subsequent warping are the results of a common stress pattern, however, the ribs and troughs would probably form perpendicular to the direction of thrust movement (axial plains perpendicular to the maximum stress or compression), and this is the case in the Silurian Hills.

Intrusive Pods

Lenses and small irregular masses of medium- or coarse-grained granitic rock, here called pods, are numerous throughout the Silurian

Hills area, occurring abundantly even where the host rock is unmetamorphosed. They are most abundant in the chaos and in areas adjacent to chaos; no pods were found that could not be interpreted as being near chaos. Pods are generally about 6 feet wide by 40 feet long but range from 1 to 300 feet in maximum dimension. They are formed from equigranular rock, generally gray—quartz monzonite or granodiorite in all cases examined carefully. Most of the pods are crushed and altered so that texture and structure are obscure. The distribution of these pods is shown in Plate 1 and Figure 8.

Pods commonly are localized along faults and other fractures, and the contacts show later movement. In some places several walls of an irregularly shaped pod suggest only minor movement, but one or two walls are faced with gouge and coincide with larger faults (Fig. 10A). A few pods have igneous contacts and chilled border phases (Fig. 10B-C). Baked contacts are rare; contact metamorphic minerals are found at only a few places and then only in limestone.

The pods are not fault slivers brought in from another area, because some of the faults along which the pods are localized have not developed breccia fragments and have gouge zones less than an inch thick. A few igneous contacts were found. It is difficult to imagine that rock of this grain size was intruded in its present position near the surface or in association with a relatively thin thrust plate that probably formed at or very near the surface. Medium-grained granitic rock is generally assumed to form by slow cooling at considerable depth. Rocks intruded near the surface generally have a chilled border phase.

The most logical explanation of the above observations is as follows: Intrusion occurred at shallow depth in an area that had a high geothermal gradient (volcanic area?). The pods were intruded into the chaos structure in the final stages of thrust movement, and in many cases partially crystallized mushlike pods were faulted and deformed. Postintrusive thrust movements sheared and rearranged the pods.

Tertiary(?) sediments and volcanic rocks were laid down in the Silurian Hills area not long before thrusting. Although these rocks were folded before thrusting (or possibly during thrusting), volcanism might still have been going on in the general area during the final stages of thrusting. Volcanism may have heated the country rock enough to allow medium-grained granitic rock to crystallize. If the intrusive masses were small, they would have been incapable of producing much contact metamorphism.

If the pods were intruded into the chaos zone in the final stages of thrusting, some of the movements would have occurred while the magma was incompletely crystallized. The movement would have been localized along these mushlike masses, as they would have been zones of weakness. The movement would have stirred up the masses and spread them out, making them even more podlike, and would have mixed up and disguised the chilled border phases that did form. The final thrust movements would have occurred after all the pods had solidified and would have faulted and crushed them still more.

FIGURE 10.—PODS OF GRANITIC ROCK

A. Typical pod of granitic rock with fault contacts. Two contacts are straight, sharp faults. The third contact is sinuous and irregular but is slickensided and gouge bearing. No chilled border is visible, but the whole mass is sheared and broken, and the border phase may have been destroyed. The country rock is buff dolomite (27400 N.–28600 W.).

B. An irregular pod of granitic rock with chilled border phase (not shown) has intruded along the contact between carbonate rock and siltstone of the Riggs formation (26400 N.–33900 W.).

C. Detail of the intrusive contact of B. Unlike most of the granitic pods in the Silurian Hills, this pod has a fine-grained border phase 1–8 inches wide.

ORIGIN OF THE CHAOS

Amargosa Chaos

Noble (1941) first described "chaos" structure in the Death Valley area. He ascribed its origin to:

> "A flat thrust of middle or late Tertiary age. . . The rocks of the overthrust plate are broken into innumerable blocks and slices, which are thrust over one another to form . . . the Amargosa chaos. . . (p. 942). The Amargosa chaos as a whole has moved out over the underlying floor. . . (p. 977). Several lines of evidence indicate the amount of movement was large and the relative movement of the overthrust plate was toward the west . . . from outcrops of what is believed to be the same thrust, its apparent minimum extent is 75 miles from northwest to southeast (p. 980). . . . the thrust faulting took place at shallower depth and under lighter load than the movements that have produced thrust faults of the more usual type (p. 978).
>
> "The tight packing of the blocks and the absence of slices of tectonic breccia in the Virgin Spring phase and part of the Calico phase indicate that they

moved under heavier load than the Jubilee phase. . . (which) may be the product of a considerably later movement. . ." (p. 978).

Noble brought remarkable order to the "chaos", and I am in essential agreement with most of his concepts. Without his pioneer work, I could not have made an intelligible map of the Silurian Hills and would not have understood its structure.

In a more recent paper, Noble and Wright (1954, p. 152) modified the original ideas in part:

"It seems probable that the movement of the Calico phase was independent of, and less than, the movement of the Virgin Spring phase and that it was separate from, though related to, the movement of the Amargosa thrust. The date of the thrusting of the Calico phase is assumed to be post-Miocene, for the rocks of this phase are those of the younger Tertiary group."

They state that the Jubilee phase is essentially a megabreccia (not a true chaos ?) and younger than the Pliocene-Pleistocene(?) Calico phase. They further suggest (p. 152) that the Amargosa thrust was produced by "the squeezing and arching of the Black Mountain wedge under intense compression." This seems to imply less over-all displacement on the Amargosa thrust than the 75 miles generally interpreted from the original paper (Noble, 1941, p. 980), and from the context and conversations with both authors I think this is their intent. This is more nearly in agreement with the estimated 5–10 miles displacement of the Riggs thrust plate.

Brecciation on a gigantic scale is generally found associated with strike-slip faults like the San Andreas fault. An interesting and instructive description by Edgar Bailey of the breccia along an old course of the San Andreas fault is soon to appear in a U. S. Geological Survey Professional Paper on the New Almaden, California, quicksilver deposit. Noble and Wright (1954, p. 157) describe the Death Valley trough as the locus of a "great strike-slip fault like the San Andreas" (Figs. 1, 2). This fault should have strongly brecciated the rocks over a wide zone, and it is interesting to speculate on the role this breccia may have played in the formation of chaos. The brecciation did not form the Amargosa chaos (or the Riggs chaos) as it is today, but pre-thrust strike-slip brecciation of the rocks might have caused some of the chaos effects. Strike-slip faulting may be the cause of the brecciation of the rocks that have since been reformed into megabreccias.

Riggs Chaos

In the Silurian Hills the Riggs chaos is not the thrust plate but occurs under the thrust plate. It is made up of material not from the plate but from the footwall. These changes from the type chaos structure of Death Valley suggest minor modifications of Noble's ideas on the origin of the chaos as applied to the Silurian Hills.

In the Silurian Hills, chaos structure probably did not form by the movement of one or a few large thrusts plates over great distances and their subsequent shattering but rather by the imbrication and piling up of a series of small thrust plates, none of which moved very far. The strongest, most competent thrust plates, like the Riggs plate, held together as a coherent mass and moved several miles without being broken appreciably. Other less competent units broke into a series of small blocks or slices that moved only a short distance and piled up in an imbricate structure. After a number of these small slices piled up together, the resulting composite plate was stronger than the individual slices and was able to move for a greater distance—possibly several miles. The incompetent material may have broken into fault breccia or gouge and acted as a lubricant for the larger competent blocks and plates.

Rocks in the earth's crust are not strong enough to transmit forces horizontally along the surface for a great distance (as a stress is transmitted along a structural beam). Thus great thrust plates cannot be thought of as having been pushed from behind. In a thrust such as the Riggs, the stress must have been transmitted upward from rocks lying well below the whole thrust area. These deep-seated rocks are best thought of as having been deformed under compression by flowage, shearing, or some similar process. As the basement shortened, the thin shell of sediments also must have shortened. Where the shell consisted of competent rocks close to the surface, deformation was by thrusting and imbrication, and chaos structure resulted. In a general way the process may be likened to wax film on a stretched rubber sheet; as the sheet contracts the wax is fractured and imbricated. The analogy would be closer from the point of view of relative strength if the film were formed of a concentrated suspension of fine sericite flakes in light oil.

Diagrammatic Presentation of Origin of Chaos

Figure 11, intended to show one possible origin for the chaos, is a diagrammatic presentation that represents observed field conditions only in a general way. In nature each block and slice in the diagram would be broken up into numerous, smaller blocks and slices, and these smallest units would be intensely brecciated and broken. The vertical scale has been exaggerated about 30 times, so that the apparent steep dips and the steep surface slopes are actually essentially horizontal. No attempt has been made to show the deformation and shortening in the underlying crystalline rocks. For simplicity, deformation is described as proceeding from right to left, as this is more nearly in accord with the observed location of megabreccias in both the Silurian Hills and Death Valley areas, but it could equally well have proceeded from left to right or in an irregular manner.

From stage 1 to stage 8 the original block, about 100 miles long, is shortened to two-thirds its original length. A very thin surface film on this block is thickened by imbrication and chaos formation to about 500–1000 feet. If a block 4½ inches long were compressed to 3 inches (Fig. 11), the unexaggerated thickness of the final chaos structure on the block would be equivalent to a coat of paint 0.01 inch thick.

STAGE 1: A block of the earth's crust is shortened. Brittle material on the surface rests on older crystalline rocks. A break occurs at some weak spot, such as c. The relative strengths and structures at this point determine in which direction the newly formed thrust surface will dip. The direction of dip will establish one block (in this case A) as moving relatively under the other (B). Block B may also be considered as moving relatively over A. The amount of crustal shortening is indicated by S_1, and the potential surfaces of movement for the next stage are indicated by dotted lines.

STAGE 2: As the rocks at depth are shortened by amount S_2, three more breaks occur in the surface material, and the thrust imbricates. Incipient breaks are shown at b and e.

STAGE 3: As new breaks form, the new thrust masses at b move forward amount S_3. They carry the older imbricated material at c forward on their backs. Meanwhile, movement also occurs at e and shortens the crust by amount S'_3. The whole section e-c-b is now a chaos.

STAGE 4: The crustal layers in area d are composed of a strong, rigid material that tends to move as a unit (like the Riggs thrust plate). As the deeper-seated rocks continue to be shortened, this block breaks loose as a large thrust plate and rides out over e-c. The surface rock shortens by amount S_4. The wedge of rock above f-e and below plate d is too tightly contained to fracture and thrust, so it folds and thickens. Plate d is arched up in riding over e-c, and the folding arches it more. The section f-e-d-c-b-a is now thickened sufficiently to act as a unit. The weight of the section "welds" it to the underlying basement, and it now acts as a buttress to further thrusting in adjacent areas.

STAGE 5: The surface rocks are shortened by amount S_5 along three breaks at g. The three blocks are piled into an oversteepened and unstable front. New breaks are about to form near g and h (*see* dots).

STAGE 6: The oversteepened front of block g slumps off or breaks loose in a giant landslide or similar gravity-controlled debris flow. The block is shortened by amount S_6, and the resultant megabreccia, like the one in the Silurian Hills, is overridden in part by the thrust masses. (If such a younger thrust is present, it is under the alluvium north of the Silurian Hills.) The section from g to h is relatively strong, and the other breaks occur at h. Three imbricate masses pile up in an oversteepened front as the block is shortened by amount S'_6.

STAGE 7: The oversteepened front slumps off in a debris flow, and the block shortens by imbrication at k in amount S_7. Continued landsliding and erosion at h weakens the crustal rock at h and prepares the way for a large thrust to form under the hill at j.

STAGE 8: As the block is shortened amount S_8, the thrust fault under the hill at j carries k-j out over h, shoving out a large amount of debris in front of it, and finally overriding it. The jumbled mass under h-j is transitional between megabreccia and chaos in general structure. Block h is similar to the Virgin Spring phase and block k to the Calico phase of the Amargosa chaos (Noble, 1941). The jumble at j is possibly similar to the Jubilee phase (Noble and Wright, 1954).

Figure 11 shows a suggested origin for chaos structure and is not intended as an explanation of thrust faulting in general. It illustrates general processes and not specific details. The shortening has been described as a series of steps, but many of these steps would be taking place at the same time.

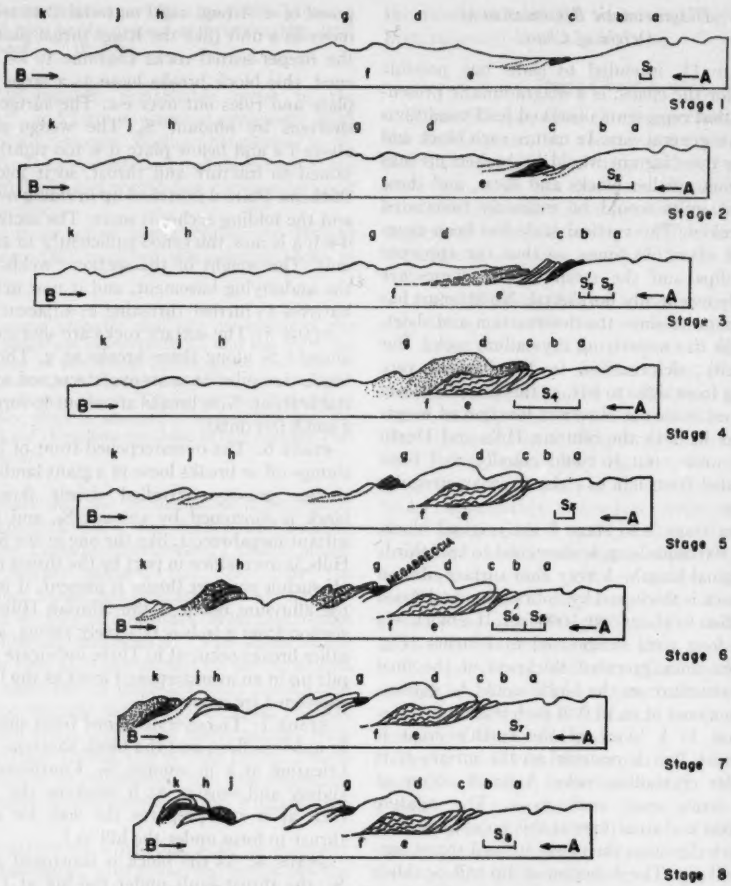


FIGURE 11.—DIAGRAMMATIC SKETCH SECTIONS SHOWING POSSIBLE ORIGIN OF CHAOS STRUCTURE BY THRUST FAULTING

The earth's crust under compressive strain from an unknown stress fails in the surface layers by imbrication. The deformation in the rocks under the surface layers is not shown. The vertical scale is exaggerated about 30 times, causing the relatively flat dips to appear steep. Sketches are simplified and diagrammatic and do not exactly correspond to known areas. Each stage shows a shortening of the earth's crust by amount S_n, but the stages are not necessarily separable or chronological. Details are explained in text by reference to the areas a, b, c, d, g, h j, and k.

The movement of plate d in stage 4 is likened to the Riggs thrust plate, but in the figure the chaos under the plate is formed first and then overridden by the plate. I believe, however, that the Riggs chaos formed under the Riggs thrust as this thrust moved forward. The arching of plate d as it moved forward is exaggerated by the great vertical exaggeration of the figure and only partly explains the folding of the Riggs thrust, if at all. Much of the folding probably took place after the thrusting.

Other Hypotheses

Sears (1953) postulated that chaos structure forms by gravity sliding. Rising magma arches up the surface of the earth, and the chaos blocks separate along "hospitable horizons" and "...do not rise up with the hill in the first place. Their relative elevation probably does not greatly change throughout the process, but, as the basement rises out from under them, they readjust themselves slowly and more or

less imperceptibly" (p. 186). He placed particular emphasis on the "widespread talc bed, which, owing to its lubricating qualitites, has in many places been the zone along which movement occurred" (p. 183).

Sears' hypothesis does not fit the chaos structures observed in the Silurian Hills for several reasons, the most important of which are: (1) The chaos material has been derived mainly from the rocks under the thrust surface, not from the rocks above it. (2) No major surface of slippage underlies the chaos structure; instead, faulting and brecciation are less intense downward and die out in the granitic and metamorphic complex. (3) Nowhere in the Silurian Hills does the talc horizon appear to be an important surface of movement.

Jahns and Engel (1950) mentioned unusual breccias throughout southern California which they stated "... are sedimentary rocks ... (that) resemble modern mud-flow and debris-flow accumulation." This interpretation may be correct, in part, for the megabreccia of the Silurian Hills, but as has already been explained, it does not seem to apply to the Riggs chaos.

Conclusion

Every gradation exists between sedimentary or debris-flow breccias and chaos structures, but they should not be confused. Typical chaos, like the Riggs chaos or the Virgin Spring chaos, has been described by Noble (1941) in nongenetic terms. It should be recognized by its field relations, not by its origin.

The tectonic nature of the Riggs chaos is clear. It involves the rocks of the underlying block to a much greater extent than it involves the rocks of the plate. It contains very little breccia material that could be interpreted as similar to that found in mud flows, debris flows, talus slopes, or landslide masses. It contains no sedimentary material of contemporaneous origin. The Riggs chaos formed from a series of imbricate thrusts of small net slip, probably as a result of crustal shortening.

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Short Notes

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SILURIAN FISH FOSSILS IN THE SALINA BASIN

BY W. P. LEUTZE

Abstract

New Silurian cyathaspid fish localities have been found in the Wills Creek formation of West Virginia and Maryland. Late Silurian fishes had been reported previously from the same states, Pennsylvania, New Jersey and New York. In the Salina basin, occurrences are restricted to the eastern margin, chiefly in or near the red clastic facies. Eurypterids of the same basin have a different distribution pattern and are found principally in argillaceous carbonates.

Denison (1956, p. 380–385) summarized the data concerning the known occurrences of Silurian fishes in eastern North America. Only six localities were cited, from New Brunswick to Maryland. Two of these are in pre-Cayugan rocks. The other four (Vernon shale of New York; High Falls shale of New York–New Jersey; Tonoloway and Bloomsburg formations of western Maryland; Bloomsburg and Landisburg sandstones of Perry County, Pennsylvania) are in strata that were deposited along the eastern margin of the Salina (Late Silurian) basin (Fig. 1).

Denison's reference to Silurian fishes in western Maryland is based on Swartz *et al.* (1923, p. 43), who stated that "fragments of fish scales" are found "in the red beds of the Bloomsburg member" (in Maryland?), but they gave no locality and did not mention these in the measured sections. On page 220 they listed *Palaeaspis americana* Claypole from the Tonoloway formation of Maryland but indicated no locality or exact stratigraphic position. In 1956, the writer found a bed of green and purple mottled argillaceous sandstone at Round Top, Maryland. This bed, 11 inches thick, contains numerous Heterostraci fragments. The latter are well-preserved black dentine plates up to 12 mm long. They are associated with unidentified ostracodes and *Lingula* cf. *L. semina* Ruedemann. The section has been described by Swartz *et al.* (1923, p. 152–162), but the fish bed was not noted. It is in the Wills Creek formation, 9 feet 9 inches to 10 feet 8 inches above the top of the Bloomsburg sandstone. Vertebrate fossils have not been reported previously in the Wills Creek formation of Maryland.

Woodward (1941, footnote, p. 200) described a sandstone lentil believed to be in the Wills Creek formation. This bed is about 60 feet above the Williamsport sandstone and "contains a thin seam full of fish scales or plates" where it crops out on State Route 28 about a mile east of Huntersville, Pocohontas County, West Virginia.

While searching for eurypterids, the writer recently discovered an exposure near Bass, Hardy County, West Virginia, where a few fish plates were collected. A narrow dirt road runs northwest from Bass up the east side of Elkhorn Mountain. About 2136 feet (paced) northwest of Bass, the road fords a stream. Hard, black, calcareous shale interbedded with irregular gray limestone lentils crops out in the stream bed 522 feet upstream from the ford. The stream flows along the strike of the beds for several yards at this place. In the limestone lentils, Heterostraci plates are rare but exceptionally well preserved. The exposure is estimated to be approximately in the middle of the Wills Creek formation, probably 100 feet stratigraphically above the top of the Williamsport sandstone which is exposed on the same stream. Thus it is not far from the fish horizon reported by Woodward in Pocahontas County.

These southern fish localities, when added to those previously reported, form a pattern. All the known occurrences are arranged in a single line along the eastern margin of the Salina basin. This observation may help in the interpretation of the habitat of these Silurian vertebrates. Some authors (Chamberlin, 1900, p. 402–404; Romer, 1933) have contended that the early fishes and eurypterids lived together in a fresh-

water environment. Denison (1956, p. 371) distinguished the Osteostraci-Anaspida-Eurypterida assemblage from the association of Heterostraci (Cyathaspinae) and Acanthodii. It is significant that Osteostraci and Anaspida have not been found in the Salina basin. Whatever the habitat of the early vertebrates, it appears improbable that Heterostraci and eurypterids of the Salina basin occupied the same environment. In the Salina basin, the pattern of eurypterid localities shows little relation to the distribution of fish (Fig. 1). Moreover, the two are normally found in different facies. Most of the fish fossils in Upper Silurian strata of this region are contained in clastic rocks and in several places are in red beds. The eurypterid remains are most often preserved in argillaceous dolomites and are unknown in red lithologies.

There are two exceptions: Eurypterids are found in the same lentil with fossil fishes (*Vernonaspis* and *Archegonaspis*) near Vernon, New York, but this association is atypical. The eurypterids are preserved as macerated fragments, and this is the only place in the Salina basin where fish fossils have been found in the same bed with a marine fauna. The latter includes cephalopods, articulate brachiopods, and a siphonophore. At Bass, West Virginia, there are eurypterids both higher and lower than the fish horizon, but the fossiliferous beds are separated stratigraphically by several tens of feet.

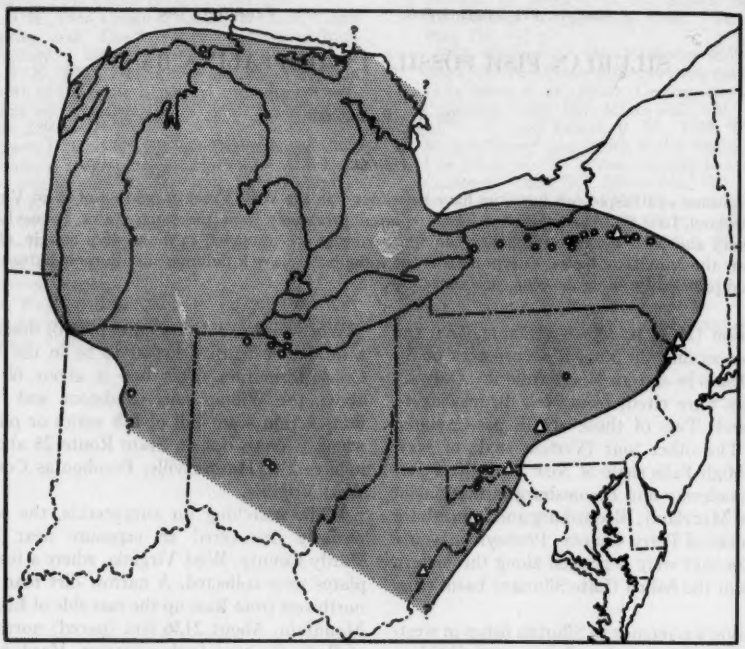

FIGURE 1.—PALEOGEOGRAPHIC MAP OF THE SALINA BASIN

Late Silurian fish localities are indicated by triangles, eurypterid localities by circles. The map represents time of deposition of the following stratigraphic units: Pointe aux Chenes (Michigan); Kokomo (Indiana); Bass Islands group (Ohio); Bloomsburg, Wills Creek, and Tonoloway (Pennsylvania, Maryland, West Virginia); Salina group (New York). The Put-in-Bay member of the Bass Islands group and the upper half of the Tonoloway may be post-Silurian.

Representative Round Top specimens have been deposited in the Ohio State University Museum (no. 22238 and 22239). The Bass specimens are in the Chicago Natural History Museum (PF 2131 and PF 2132).

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POTASSIUM-ARGON AGES IN THE PRECAMBRIAN BASEMENT OF COLORADO

BY C. E. GIFFIN AND J. L. KULP

Earlier geochronometric work in the Rocky Mountain region has established the presence of ancient basement rock (2600 m.y.) in the vicinity of the Beartooth and Bighorn uplifts in southern Montana and northern Wyoming (Gast, Kulp, and Long, 1958). Orogenic effects over a considerable area from the Laramie Range of southern Wyoming through Colorado and New Mexico into Arizona appear to have occurred in the range of 1300–1400 m.y. ago (Aldrich, Davis, Tilton, and Wetherill, 1956; Aldrich, Wetherill, Davis, and Tilton, 1958). Measurements on mica and zircon from granite in the Pikes Peak area (Tilton, Davis, Wetherill, and Aldrich, 1957) indicate a still younger event which occurred in this region at about 1000 m.y. ago. Hutchinson (1959) has recently reported several K-A ages on rocks from the area of the Pikes Peak batholith.

The purpose of this study was to survey further the Precambrian rocks of Colorado with two major objectives in mind: (1) to search for possible occurrences of the 2600 m.y. old basement, and (2) to delimit the areas affected by the 1400 and 1000 m.y. events.

The age measurements were made on separated mica by the potassium-argon method using the isotope-dilution technique as described by Long and Kulp (in press). The

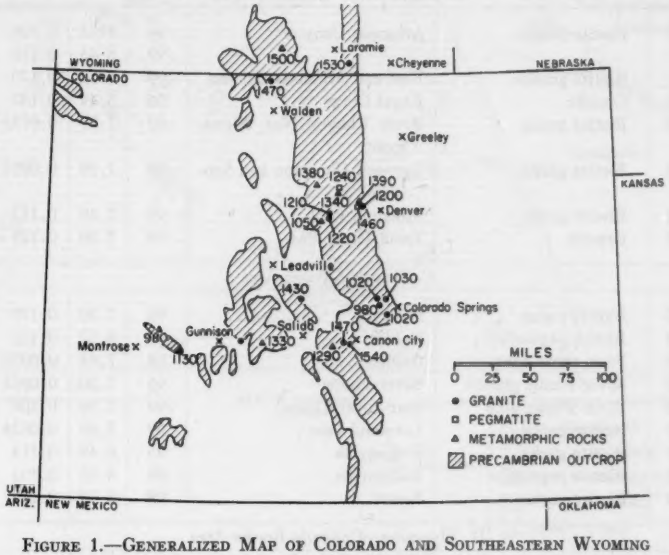

FIGURE 1.—GENERALIZED MAP OF COLORADO AND SOUTHEASTERN WYOMING

A^{40}/K^{40} ratio is believed to be accurate to at least ±4 per cent. The decay constants for K^{40} used in calculating the results are $\lambda_\beta = 4.72 \times 10^{-10}$ yr.$^{-1}$ and $\lambda_e = 0.584 \times 10^{-10}$ yr.$^{-1}$. All samples appeared free from obvious weathering effects, but some contained appreciable chlorite. Although in most cases the presence of chlorite does not appear to affect the A^{40}/K^{40} ratio, it seems possible that in extreme cases some loss of radiogenic argon may occur. The maximum concentration (18 per cent) of chlorite in these samples was observed in the biotite from the gneiss at Loveland Pass (sample PUF-21B). The locations are shown on the generalized map of Colorado and southeastern Wyoming (Fig. 1). The analytical results are compiled in Table 1.

The isotopic ages are rounded off to the nearest 10 m.y. The analytical error is ±40 to ±60 m.y. in the age range 1000 to 1500 m.y., respectively. These ages will represent the time

TABLE 1.—POTASSIUM-ARGON AGES FROM THE COLORADO BASEMENT

Sample no.	Type of rock	Locality**	%A^{40*}	%K	A^{40*}/K^{40}	Isotopic age (m.y.)
		A. Pikes Peak Area				
PUF-9B	Windy Point granite	Top Pikes Peak	89	6.00	0.0751	980
PUF-10B	Pikes Peak granite	Ute Pass	93	5.32	0.0795	1030
PUF-16B	Mount Rosa granite	Gold Camp Road	80	6.86	0.0790	1020
PUF-17B	Pikes Peak granite	Pikes Peak Toll Road	97	7.11	0.0775	1010
			91	7.11	0.0803	1030
		B. Canon City—Gunnison Area				
PUF-7M	Biotite gneiss	Arkansas Canyon	90	8.43	0.108	1290
			99	8.43	0.110	1300
PUF-8B	Biotite granite	East end Arkansas Canyon	99	6.22	0.130	1470
PUF-29B	Granite	Royal Gorge	96	5.49	0.139	1540
PUF-13B	Biotite gneiss	Black Canyon Nat. Monument	62	7.09	0.0753	980
PUF-14B	Biotite gneiss	Between Cimarron and Sapinero	98	7.59	0.0915	1130
PUF-34B	Biotite gneiss	Monarch Pass	99	7.40	0.113	1330
PUF-24B	Granite	Trout Creek Pass	98	7.10	0.125	1430
		C. Denver—Fraser Area				
PUF-32B	Biotite gneiss	Golden	99	7.20	0.120	1390
PUF-35B	Biotite pegmatite	Golden	97	6.52	0.128	1460
PUF-37B	From shear zone	Golden	98	7.64	0.0976	1200
PUF-19B	Silver Plume granite	Silver Plume	96	7.70	0.0993	1210
PUF-23B	Silver Plume dike	near Silver Plume	99	7.16	0.100	1220
PUF-21B	Biotite gneiss	Loveland Pass	99	5.89	0.0824	1050
PUF-22B	Biotite gneiss	Rollinsville	95	6.49	0.114	1340
PUF-26B	Biotite pegmatite	Rollinsville	69	8.08	0.102	1240
PUF-18B	Biotite gneiss	Fraser	98	5.79	0.119	1380
		D. Wyoming—Colorado Border Area				
PUF-30B	Granite	near Northgate	86	7.41	0.0953	1180
			98	7.41	0.0936	1160
PUF-25B	Granite	Buford, Wyoming	97	5.64	0.138	1530
			98	5.64	0.139	1540
PUF-27B	Biotite gneiss	West of Centennial, Wyoming	97	7.69	0.133	1500

* Radiogenic component
** Colorado unless indicated

of the last metamorphic or igneous event in the area, unless the rock lies in a transition zone where an older rock has been reheated by a younger thermal event which has caused partial loss of the accumulated radiogenic argon (Long, Kulp, and Eckelmann, 1959; Long and Kulp, in press).

The data (Fig. 1; Table 1) give no evidence of a basement older than about 1500 m.y. If a 2600 m.y. basement complex did exist in these areas of Colorado the micas have been sufficiently reheated by an event in the period 1300–1500 m.y. ago to have lost all their inherited argon. Further studies using the Rb-Sr method on microcline and the U-Pb method on zircons will be required to detect the presence of older material.

The youngest metamorphic-igneous event in the Colorado basement appears to have occurred at about 1000 m.y. ago. It seems thta this activity centered in the Pikes Peak region and may have produced some of the granites

in the vicinity such as the Windy Point and the Mount Rosa as well as the pegmatites near St. Peters Dome. Figure 2 is a geologic sketch map of the area around Pikes Peak showing the various igneous and metamorphic units and the age determinations including those of Aldrich *et al.* (1958). The average of all the K-A determinations is 1015 m.y. with a standard deviation of about ±40 m.y. (*i.e.*, no more than the experimental uncertainty). Hutchinson (1959) reports a K-A age of 1050 m.y. for a hornblende granodiorite on the northwest side of the Pikes Peak batholith. Two Rb-Sr determinations (Aldrich *et al.*, 1958) were 960 and 1020 m.y. assuming the decay constant of $\lambda_\beta = 1.47 \times 10^{-11}$ yr.$^{-1}$ (Flynn and Glendenin, in press) in good agreement with the K-A determinations. U-Pb isotopic ages on the zircon from the lower Pikes Peak granite by Tilton *et al.* (1957) were strongly discordant, *ie.*, U^{238}-$Pb^{206} = 624 \pm 25$, U^{235}-$Pb^{207} = 707 \pm 20$, and Pb^{207}-$Pb^{206} = 980 \pm 40$.

It is possible that the event at about 1000 m.y. ago was a minor one involving only the reheating of the micas with consequent loss of radiogenic argon and strontium. On this hypothesis, the granites of the Pikes Peak region may have been considerably older. Some support for this concept can be found in the age obtained for the granite at Trout Creek Pass (1430 ± 60 m.y.) which appears petrologically identical to the Pikes Peak granite. If this correlation is valid, the granite in the vicinity of Pikes Peak may also have been emplaced at about 1430 m.y., and the presently observed isotopic age (~1000 m.y.) merely indicates the time of reheating. The other granites in this area may represent different phases or pulses of the Pikes Peak granite in which case they may also have been emplaced earlier. Further support for the hypothesis that the Pikes Peak batholith was emplaced earlier than the 1000 m.y. event is the K-A age of 1240 m.y. reported by Hutchinson (1959) on the Turkey Ridge pluton at the north end of the batholith.

The 1000 m.y. event was not restricted to the vicinity of Pikes Peak. It appears to have af-

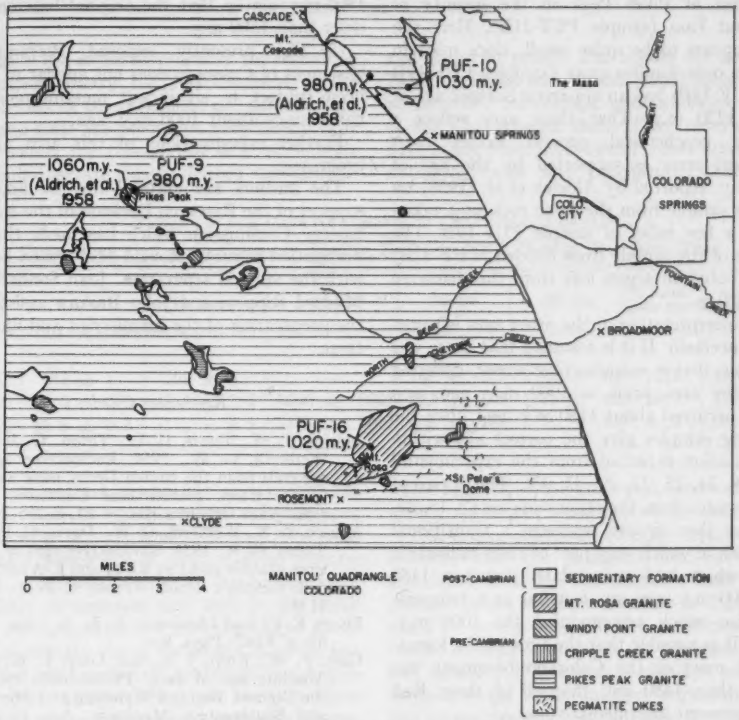

FIGURE 2.—GEOLOGIC SKETCH OF THE PIKES PEAK AREA

fected part of the basement west of Gunnison (samples PUF-13B and PUF-14B; *see* Fig. 1 and Table 1) in the Black Canyon region. It may also have affected the rocks 70 miles northwest of Pikes Peak in the vicinity of Loveland Pass (sample PUF-21B). Here the area appears to be quite small, since mica in granites only 8 miles away (samples PUF-19B and PUF-23B) has an apparent isotopic age of about 1220 m.y. That these ages reflect a genuine geochemical process rather than analytical error is supported by the age of 1230 m.y. reported by Aldrich *et al.* (1958) on another sample from the same rock unit taken within a few miles of sample PUF-19B. The low age of the sample from Golden (PUF-37B) is attributed to argon loss from the distorted mica lattice.

The interpretation of the other ages is somewhat uncertain. If it is assumed that there was only one major metamorphic event covering the entire area prior to 1000 m.y. ago, and that it occurred about 1480 m.y. ago, then the following samples give the correct age within the variation expected from the experimental error (8, 24, 25, 27, 29, 32, 35). The apparent isotopic ages from the other samples (7, 18, 22, 34) may then simply represent a transitional condition in which only part of their radiogenic argon which had accumulated between 1480 and 1000 m.y. ago was lost due to a temperature rise which accompanied the 1000 m.y. event. It is possible that the time of the formation of most of the Colorado basement was earlier than 1480 and that all of these K-A ages represent intermediate ones.

From this survey the following conclusions are reached:

(1) Further reconnaissance of the Precambrian basement of Colorado by the potassium-argon method has not detected rocks older than 1500 m.y.

(2) A major orogenic event may have occurred about 1500 m.y. ago in the area of Colorado which either produced the initial basement or recrystallized it to such an extent that no inherited radiogenic argon was preserved.

(3) A less intense metamorphic-igneous event occurred at about 1000 m.y. ago which caused loss of inherited radiogenic argon in the granites and metamorphic rocks of the Pikes Peak area and many have been accompanied by the intrusion of the St. Peters Dome pegmatite swarm. This event is clearly detected in small areas as far west as the Black Canyon of the Gunnison and north to the Wyoming border near Northgate, Colorado.

(4) All K-A ages in Colorado can be most simply accounted for by assuming only these two events. The data, however, do not preclude additional thermal pulses between 1000 and 1500 m.y. ago or that the original basement is older than 1500 m.y.

(5) The presently exposed Precambrian basement of Colorado does not appear to have been subject to significant metamorphic activity since about 1000 m.y. ago.

Further investigations of this area are in progress.

The authors are grateful for the financial support of the Research Division of the Atomic Energy Commission which has made this investigation possible. Ronald McPherson helped with the mineral separation. Joan Sonderburg, Mildred Rippey, and John Brokaw assisted in the preparation of the manuscript and illustrations.

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U-Pb AGE OF THE CHATTANOOGA SHALE

J. C. COBB* AND J. L. KULP

Despite the major advances in the technique of geochronometry the establishment of the absolute age of the geological, *i.e.*, paleontological, time scale has proceeded very slowly. The chief reason for this is the lack of suitable materials that are closely defined stratigraphically and yet are susceptible to accurate age determination by isotopic methods. Bituminous shales with a sufficiently high U/Pb ratio present a possibility of meeting these requirements. A detailed study of the isotopic geochemistry of uranium and lead in the black shale and associated kolm of the Upper Cambrian of Sweden (Cobb and Kulp, in press) has provided the necessary analytical techniques and some understanding of the alteration processes of the uranium-lead system in bituminous shales. It was therefore decided to attempt to date the upper black unit of the Gassaway member of the Chattanooga shale which carries the maximum U/Pb ratio. The upper part of the Chattanooga shale is well defined stratigraphically as uppermost Devonian (Hass, 1956).

Thanks to the generous co-operation of Professor T. F. Bates of the Pennsylvania State University it was possible to obtain samples from drill core YB-9 in the Youngs Bend area, Tennessee. The black shale unit selected for study is 4.9 feet thick in this core and carries an average uranium content of .80 ppm (Kehn 1955).

Three samples were analyzed by isotopic dilution techniques described elsewhere (Cobb and Kulp, in press). The results are given in Table 1. Bates and Strahl (1957) have made an intensive mineralogic and petrologic study of the Chattanooga shale. They conclude, as have many other investigators, that uranium is authigenic and was precipitated from sea water because of the anaerobic conditions prevailing during the formation of the shale.

TABLE 1.—AGE DATA ON SAMPLES OF CHATTANOOGA SHALE

Upper Black Unit of the Gassaway member, Drill Core YB-9, Youngs Bend Area, Tenn.

Depth (feet)	U (ppm)	Pb (ppm)	204/206	207/206	208/206	% Radiogenic Pb^{206}	Apparent U^{238}-Pb^{206} age
							m.y.
175.12	112 ± 1	192 ± 6	—	.768 ± .016	1.84 ± .04	13.2	450 ± 70
176.95	90.6 ± 0.9	47.4 ± 1.0	.0390 ± .0005	.622 ± .005	1.51 ± .02	28.7	350 ± 15
180.00	68.3 ± 0.7	31.8 ± 0.5	.0375 ± .0005	.607 ± .006	1.46 ± .01	31.4	350 ± 12

This origin insures an effective separation of uranium from the radiogenic lead previously associated with it. The common lead incorporated in the shale at the time of deposition could be partly detrital. The most important carrier of such lead would be feldspar. Bates and Strahl (1957) estimate about 9 per cent feldspar in the shale, but they regard much of this as authigenic. It is probable that most of the original lead was also precipitated from sea water, and therefore the isotopic composition of this lead can be estimated to be that of oceanic lead in late Devonian time, *i.e.*, $Pb^{206}/Pb^{204} = 18.3$; $Pb^{207}/Pb^{204} = 15.7$; $Pb^{208}/Pb^{204} = 38.6$. This is obtained by extrapolating backward from the isotopic composition of modern oceanic lead (Chow and McKinney, 1958), assuming the average U/Pb ratio in crustal rocks did not change during this relatively small fraction of geologic history. The correction for the original lead is so large that the resulting error on U^{235}-Pb^{207} age precludes its usefulness. The error on the U^{238}-Pb^{206} ages from analytical procedures and the initial lead correction are sufficiently small to warrant a serious discussion of the absolute age.

The U^{238}-Pb^{206} ages on all three samples agree within their experimental errors (Table 1).

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The average of the two samples with reasonably small errors gives an apparent age of 350 ± 10 m.y. It is now necessary to evaluate the relationship of this apparent age to the true age of the formation.

The study of the Swedish black shale (Cobb and Kulp, in press) showed that alteration in this type of environment, if present, normally consists of bulk lead loss and/or preferential loss of lead 206. In the case of the Swedish black shale excess Pb^{206} was detected, producing high U^{238}-Pb^{206} ages. This was due to the movement of Pb^{206} from the kolm, which are uranium "hot spots", into the shale. Such centers of high uranium concentration do not appear to be present in the Chattanooga shale, so that this effect should be negligible. Therefore, at worst, the U^{238}-Pb^{206} age obtained on the Chattanooga should be a reliable *minimum* age.

Radium measurements on these samples of the Chattanooga shale show that this element is now in equilibrium with the uranium in the sample. This suggests that either alteration is negligible or, if the radium (and possibly lead) is being moved about, the volume of rock having reasonably uniform uranium content is sufficiently large that no net gains or losses of lead occur. It is quite possible therefore, that the true age of the Chattanooga shale is about 350 m.y.

These data would appear to set a firm minimum age for the Mississippian–Devonian boundary of 350 m.y. and suggest that the true age of this boundary cannot be much greater than this. This is consistent with the time scale proposed by Kulp (1959) based largely on the age of biotites of granites that intrude stratigraphically well-defined rocks. The common lead to radiogenic lead ratios obtained on these samples is illustrative of the limitations of this method of dating bituminous shales. As this ratio becomes more unfavorable, the correction for original lead becomes ever more important. Normal bituminous shales, with a uranium enrichment up to 10 or 20 ppm, would not give enough radiogenic lead for an age analysis, assuming that the original lead content is the same as in the Chattanooga and Swedish shales.

Acknowledgments

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POTASSIUM-ARGON AGES IN NORTHERN MANITOBA

BY J. M. MOORE, JR., S. R. HART, C. C. BARNETT, AND P. M. HURLEY

Introduction

The dating of igneous and metamorphic rocks by radioactive decay methods has provided a quantitative basis for defining orogenic provinces within the Canadian Shield. The boundary between the Churchill (Athabasca) and Superior geologic provinces is believed to lie within the province of Manitoba and has hitherto been placed, on topographic and geologic grounds, along the Nelson River valley. It was the purpose of this investigation to test this hypothesis by age measurements, as well as to test the homogeneity in age of the crystalline rocks in the northern Manitoba part of the Churchill province. As most of this area is accessible only by air, the collection of samples for this work has been severely limited. Opportunities were afforded by the field programs of the Geological Survey of Canada and the Manitoba Mines Branch, on which Barnett and Moore have been employed, and by mineral investigations in the area by Hurley. Because other such opportunities will probably not arise for members of this laboratory group again, the results of this limited investigation are being reported now, even though they are minimal.

Precambrian Division in Manitoba

A division of the Precambrian of Canada into provinces, on a geographical or geological basis, was first presented by M. E. Wilson (1941) and elaborated by J. E. Gill (1949) and J. T. Wilson (1949; 1952). This division has been made on the basis of distinctive lithology, structural trends, and dominant types of mineralization characteristic of each subdivision, and, more recently, on gravity data and radioactive age measurements. The boundary between the Churchill province on the northwest and the Superior province on the southeast has been placed, in Manitoba, approximately coincident with the Nelson River valley. Wilson (1949, p. 235) states: "This valley was chosen because it is pre-Ordovician in age, paralleled by disturbances in gravity, and marked by changes in rock type, in direction of foliation, and in dominant types of mineralization." Wilson suggests that the Shield provinces represent primary orogenic belts.

Radioactive age measurements prove a chronologic distinction between the two provinces. U-Pb ages of uraninite and pitchblende from Saskatchewan and the Northwest Territories are in the range 1700 to 1900 million years (Cumming *et al.*, 1955; Eckelmann and Kulp, 1956), while U-Pb and Rb-Sr ages of uranium minerals and micas in pegmatite and granite in southeastern Manitoba and northwestern Ontario are in the range 2400 to 2700 million years (Collins *et al.*, 1952; 1954; Davis and Aldrich, 1953; Eckelmann and Gast, 1957; Kulp, 1957). Also, several K-Ar ages in the western part of the Superior province in Ontario have been reported by this laboratory (Hurley *et al.*, 1958) showing a range from 2400 to 2700 million years (Fig. 1).

Sample Description and Geology

1. *Oxford Lake-Knee Lake Area* (Wright, 1931). The Hayes River volcanic group, predominantly andesite, basalt, and related pyroclastic rocks, with minor clastic sediments, is overlain unconformably by the Oxford group of clastic sediments. Gneissic granite (3808) intrudes both Hayes River and Oxford groups. The "younger" granite (3809) is a pluglike body about 8 miles across, intrusive into all the above rocks.

2. *Partridge Crop Lake Area* (Dawson, 1952). Metasedimentary rocks, including carbonate members, and the Assean Lake series of andesite and basalt, with interbedded chert and clastic rocks, are the oldest observed rocks. These are intruded successively by mafic and serpentinized ultramafic rocks, granite and related pegmatite, and younger ultramafic and mafic dikes. Sample 3810 (south of Mystery and Moak Lakes on map) is from a 2- to 3-foot pegmatite dike cutting mafic gneiss. The sample location is 20 miles south of a major zone of thrust faulting and serpentinite intrusives, parallel to a belt of gravity disturbances extending across the province (J. F. Davies, personal communication).

3. *Lynn Lake Area.* The Wasekwan series of volcanic and pyroclastic rocks and clastic sediments is cut by intrusives ranging in composition from gabbro, with minor ultramafic rocks, to granite. The volcanic series is overlain unconformably by the Sickle series of conglomerate, arkose, and sandstone (Bateman, 1945). There is evidence that some granite is pre-Sickle, but dioritic and granitic rocks also cut the complexly folded Sickle series (Milligan, in press). Sample 3812 is a massive pink granite which is a late member of a complex body, mostly gneissic granodiorite, cutting the Wasekwan series; 3875 is a sedimentary gneiss of the Wasekwan series, and 3876 is from a massive granitic plug intruding Wasekwan rocks.

4. *Northern Indian Lake* (Churchill River drainage). Granite, 3907, locally foliated or porphyritic, and gneissic granodiorite, 3915, predominate. Sedimentary quartz-biotite gneiss, 3910, cut by pegmatite, also occurs in the area. (Reconnaissance mapping by the Geological Survey of Canada is in progress.)

5. *Herb Lake-Snow Lake Area* (Armstrong, 1939; Russell, 1957). Sample 3295, a drill core of quartz-biotite gneiss, was obtained in the vicinity of Miller (Stall) Lake. It is part of a sequence consisting of quartz-feldspar gneisses and granitic rocks, thought to be of sedimentary origin (Russell, 1957), and mafic volcanic rocks, including minor tuffs and diorite sills. These rocks are the oldest in the area, in part equivalent to the Amisk Series. They are overlain successively by a dominantly volcanic series and a clastic sedimentary series. The stratigraphic position of mafic and felsic intrusives occurring in the area is not definitely determined.

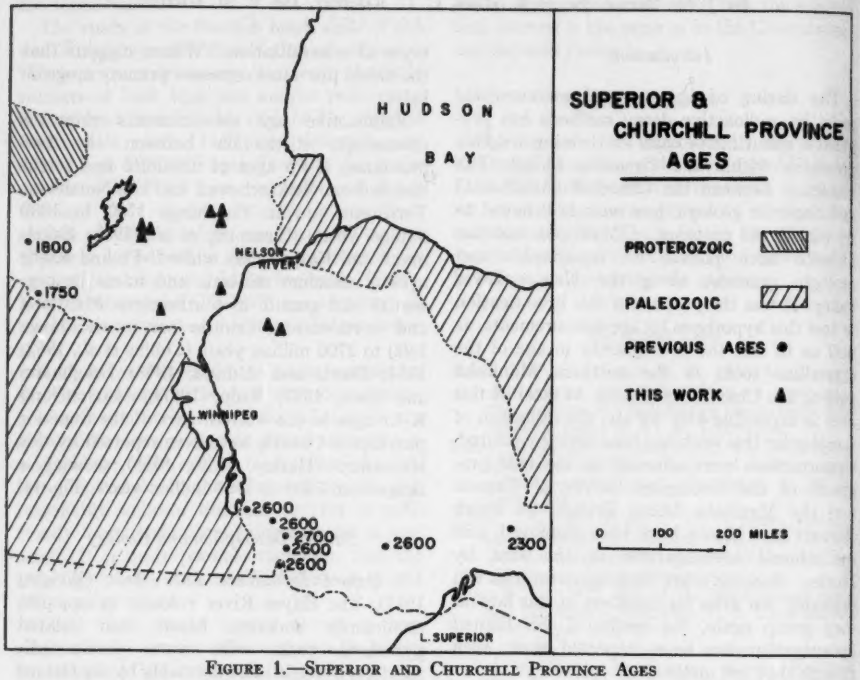

FIGURE 1.—SUPERIOR AND CHURCHILL PROVINCE AGES

Analytical Techniques

The specific techniques of age measurement used in this laboratory have been described in our Annual Reports.

The argon analyses were made by isotope dilution, using enriched Ar^{38} as a spike, and Ar^{36} was used to calculate the normal (air) argon correction. Potassium analyses were made using a Perkin-Elmer flame photometer,

with lithium internal standard. Reproducibility of argon analyses is normally better than 1 per cent. Analysis of interlaboratory standard samples, and the argon content of atmospheric

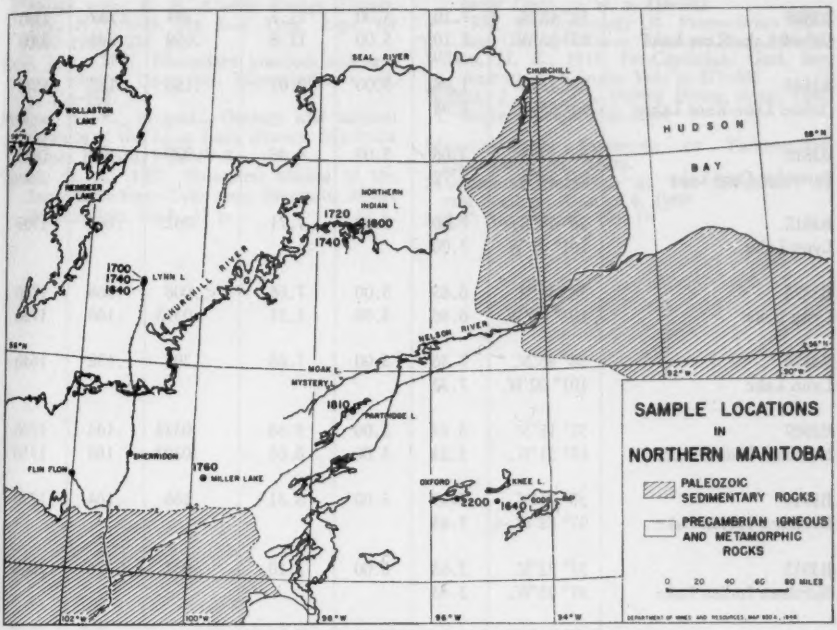

FIGURE 2.—SAMPLE LOCATIONS IN NORTHERN MANITOBA

those areas of Saskatchewan and Northwest Territories previously mentioned.

The ages from the Oxford Lake-Knee Lake are typical of "mixed-age" areas which result from the complete or partial recrystallization of older rocks in the vicinity of a younger orogenic belt. Similar inhomogeneities are observable in the Superior province where it is bordered by the Grenville province.

The central part of a block like the Superior province appears homogeneous in age because the measurements represent the time at which the present surface rocks were elevated to a temperature-depth zone in which significant loss of argon from the minerals ceased.

A younger bordering orogeny will expose rocks from depth in which low-grade metamorphism of varying degrees has developed a heterogeneity that may be broad or very local in scale. The sample B-3808, showing an age of 2200 million years, is interpreted as indicating that the normal host rock in the Oxford Lake-Knee Lake area should be included in the Superior province, although it does not show the usual 2400 to 2600 million years age.

air indicate an absolute error of less than 1 per cent. Each potassium measurement is the average of two flame photometer determinations.

The reproducibility of these potassium measurements has a standard deviation of 3 per cent. The analytical error associated with the given ages, therefore, is approximately 3 per cent. Decay constants used are $\lambda_e = .585 \times 10^{-10}$ yr.$^{-1}$, $\lambda = 5.30 \times 10^{-10}$ yr.$^{-1}$, $K^{40} = 1.22 \times 10^{-4}$ g/g of K.

Results and Discussion

The analytical details of the age measurements on biotite from the above samples are listed in Table 1 and shown on the generalized geologic map (Fig. 2).

The results from Lynn Lake, Northern Indian Lake, Miller Lake, and Partridge Crop Lake suggest that the Churchill province is homogeneous in northern Manitoba, as in

TABLE 1.—ANALYTICAL DATA FOR K-Ar AGES OF BIOTITE, NORTHERN MANITOBA

Sample no. Locality	Lat. Long.	K Per cent	Wt. of Sample (gms.)	Radiogenic Ar^{40} (cc. STP/g $\times 10^4$)	Atmospheric Ar^{40} (cc. STP/g $\times 10^4$)	Ar^{40}/K^{40}	Age (m.y.)
B3808	54° 48′N.	7.10	5.00	11.7	.234	.242	2190
Oxford Lake-Knee Lake	95° 35′W.	7.10	5.00	11.8	.059	.244	2200
B3809	54° 45′N.	1.98	5.00	2.07	.186	.153	1640
Oxford Lake-Knee Lake	94° 55′W.	1.98					
B3810	55° 39′N.	3.66	5.00	4.48	.020	.178	1810
Partridge Crop Lake	97° 20′W.	3.72					
B3812	56° 48′N.	7.00	5.00	7.71	.0625	.162	1700
Lynn Lake	101° 01′W.	7.00					
B3875	56° 46′N.	6.65	5.00	7.66	<.008	.168	1750
Lynn Lake	101° 00′W.	6.66	5.00	7.51	.0375	.166	1730
B3876	56° 46′N.	7.36	5.00	7.63	.298	.152	1640
Lynn Lake	101° 02′W.	7.32					
B3907	57° 15′N.	5.22	5.00	5.86	.0328	.164	1720
Northern Indian Lake	97° 31′W.	5.24	5.00	6.05	.0302	.169	1750
B3910	57° 21′N.	7.39	5.00	8.31	.166	.164	1720
Northern Indian Lake	97° 52′W.	7.43					
B3915	57° 22′N.	2.63	5.00	3.10	.031	.176	1800
Northern Indian Lake	97° 25′W.	2.55					
B3295	54° 51′N.	6.09	5.00	7.04		.170	1760
Miller Lake	99° 56′W.	6.09					

Acknowledgments

The writers thank Mr. J. F. Davies, Chief Geologist of the Manitoba Mines Branch, Geological Division, for providing specimens from the Partridge Crop Lake and Oxford Lake-Knee Lake areas, and for reading the manuscript, and also Mr. A. Koffman of the Hudson Bay Mining and Smelting Company for the sample of drill core from the Snow Lake-Herb Lake area. The continued support of this laboratory by the Division of Research of the Atomic Energy Commission is also gratefully acknowledged.

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APPARENT-DIP COMPUTER

By Lowell R. Satin

J. B. Jukes (1859) compiled correction tables for dip in directions not perpendicular to strike (apparent dip). A number of methods for computing apparent dip have been published since 1859, but only three methods are generally used: (1) dip-correction tables (Lahee, 1952), (2) alignment diagrams (Palmer, 1918), (3) the dip-correction protractor (Smith, 1925).

The construction of tables and mechanical aids to compute apparent dip is based on the trigonometric relationship between the true dip (α), the apparent dip (γ), and the acute angle between the strike of the bed and the desired vertical section, which is called in this paper the "convergence angle (β)."

$$\tan \gamma = \tan \alpha \sin \beta \qquad (1)$$

Equation (1) is derived in the following manner: In Figure 1,

α = true-dip angle
β = convergence angle
γ = apparent-dip angle
AB = strike of the bed
$ABCD$ = plane of the bed
$BGCH$ = vertical section normal to the strike of the bed
$BEFH$ = vertical section at an angle β to the strike of the bed

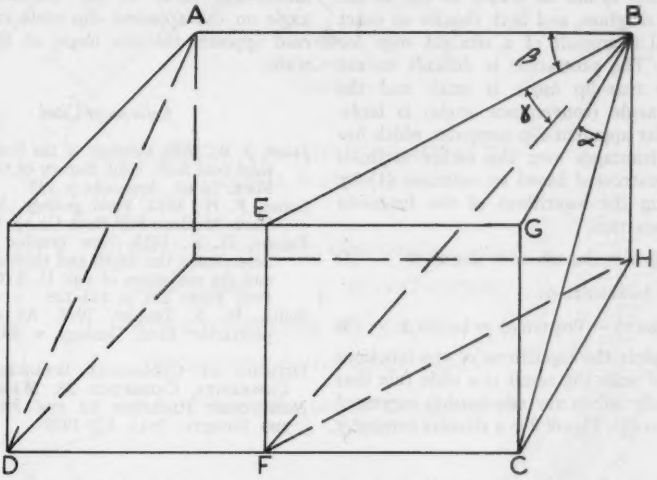

Figure 1.—Block Diagram Showing the Relationship between the True Dip (α), the Apparent Dip (γ), and the Convergence Angle (β)

$$BG = BE \sin \beta \quad (\angle ABE = \angle BEG) \qquad (2)$$
$$GC = BG \tan \alpha \qquad (3)$$
$$EF = BE \tan \gamma \qquad (4)$$

Substituting equation (2) for BG in equation (3)

$$GC = BE \sin \beta \tan \alpha \qquad (5)$$

But $EF = GC$ (*see* Figure 1). Substituting equation (4) into equation (5)

$$BE \tan \gamma = BE \sin \beta \tan \alpha \qquad (6)$$

or eliminating BE

$$\tan \gamma = \sin \beta \tan \alpha. \qquad (1)$$

Tables for the correction of dip can be compiled by using equation (1), and the results can be expressed to the nearest minute of arc (Lahee, 1952). To be accurate the tables should be large enough to include corrected values for each degree of dip and convergence angle (8000 entries or more), and they are therefore too large for convenient use. As a result, most tables are designed in 5-degree

increments, necessitating interpolation for the intermediate values, with a resulting loss of accuracy.

Palmer's (1918) alignment diagram for computing projected dip (apparent dip) is easy to use but lacks the precision of the dip-correction tables. To be accurate the diagram must be greatly enlarged when the angle of projection (convergence angle) is greater than 50°.

The dip-correction protractor designed by Smith (1925) is not as simple to use as the alignment diagram, and both require an exact mechanical alignment of a straight edge for operation. The protractor is difficult to use when the true-dip angle is small and the direction angle (convergence angle) is large.

A circular apparent-dip computer which has several advantages over the earlier methods can be constructed based on equation (1) by substituting the logarithms of the functions into the equation.

$$\log \tan \gamma = (\log \tan \alpha) + (\log \sin \beta) \qquad (7)$$

which can be rewritten:

$$(\log \tan \gamma) - (\log \tan \alpha) = \log \sin \beta \qquad (8)$$

If one plots the logarithms of the functions on a linear scale the result is a slide rule that mechanically solves the relationship expressed in equation (8). Figure 2 is a circular computer based on the slide-rule principle. The directions for using the computer are printed on the dial face.

The circular apparent-dip computer shown in Figure 2 is simple to use, accurate, and compact. A single setting of the convergence index opposite the convergence angle computes the value of any apparent dip for a given true dip, and vice versa. The computer is also useful as a drafting aid. Any desired projection angle can be set up over the convergence-angle index; the value of any desired projected angle on the apparent dip scale can then be read opposite the true angle on the true dip scale.

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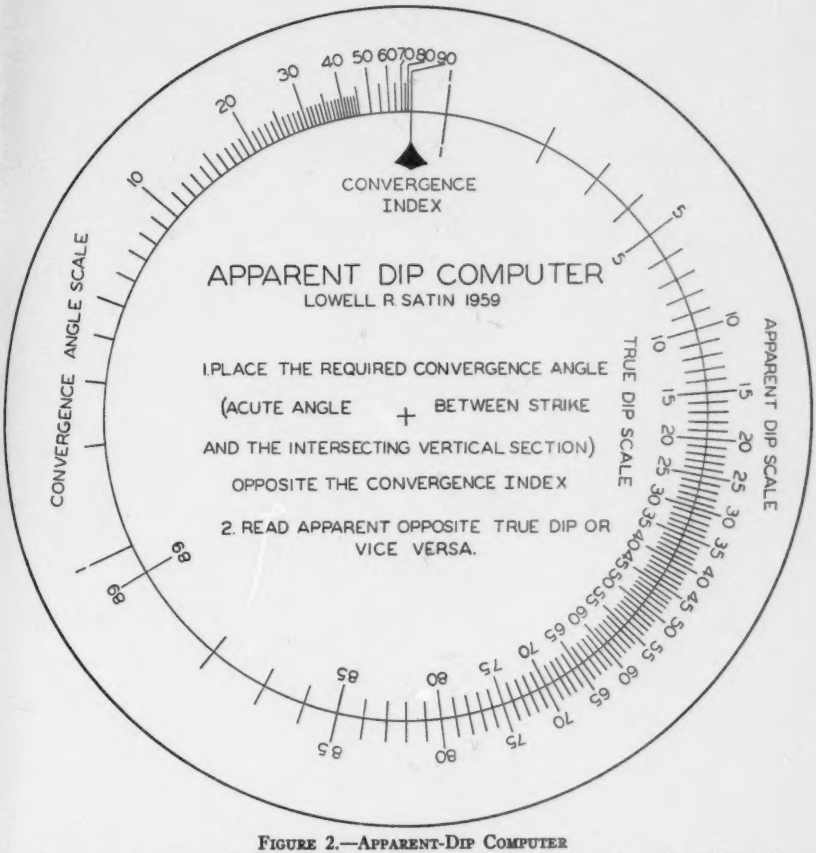

FIGURE 2.—APPARENT-DIP COMPUTER

Cut out computer on the circular line between the true dip and apparent dip scales. Mount the outer ring on a stiff backing and attach the inner dial at the center marked position. Instructions for use are printed on the dial face.

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